Fluid Migration at Opouawe Bank, offshore New Zealand

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Abstract

Fluid migration and its near-surface expression is a common and wide-spread phenomenon on subduction zone accretionary margins. Understanding the nature, variability, and lateral distribution of fluid migration has implications for the global carbon cycle, geo-hazards (e.g. seafloor stability), and hydrocarbon exploration. Additionally, insights can be gained into the spatial and temporal processes associated with the overall subduction system.

In this framework, a cooperative study between New Zealand and Germany was launched to investigate several areas of known gas seepage off New Zealand on the Hikurangi margin. During the NEMESYS project, surveys were conducted using the R/V Sonne to gather geophysical, geochemical, and microbiological data around several vent sites, including Opouawe Bank. This study utilizes mostly geophysical data sets (2D, 3D seismic and hydro-acoustic surveys), with some additional constraints from pore-water geochemical analyses.

Three case studies highlight the different aspects of the gas migration systems from the gas source to the seabed. The first study links 2D multi-channel seismic data, seismic velocity analysis, geochemical, and isotopic pore water compositions to investigate gas migration processes. Key findings include that the seepage is driven by biogenic gas migration without a substantial water phase (i.e. a "dry system"). Flow appears mostly focused into vertical conduits with some stratigraphically controlled flow along dipping horizons. The gas is generated at a maximum depth of 1,500 – 2,100 meter below seafloor (mbsf).

The use of the P-Cable seismic system allowed 3D seismic imaging and spatial analyses of the vent systems ultimately linking these systems into the tectonic framework and overall stress regime of the accretionary prism. Migration of gas was found to be along extensional fractures systems that are elongated perpendicular to the ridge axis of Opouawe Bank. Thus, these structures are parallel to the main direction of compression induced by subduction and are the result of secondary extensional forces from the gravitational collapse of the ridge. With the reduction of overburden stress the extensional structures develop into more pipe-like conduits in the shallow sediment at < 100 mbsf.

The shallowest expression of gas migration was investigated using hydro-acoustic data and 2D seismic data in the third case study. Clear differences in gas migration structures suggest a progression through different evolutionary stages of seep formation. These range from channeled gas

flow, to gas trapping beneath relatively low-permeable horizons, and overpressure accumulation associated with doming. Limits in the required height of the gas column were estimated using the spatially constrained geometry of the individual doming structure and assuming representative mechanical properties of the sediments. Contrary to ongoing discussions, this study demonstrated that these doming features can evolve into seep sites (associated with seafloor chemosynthetic communities) without the formation of pockmarks.

The combined results of these case studies suggest that the exceptional observation of extensional gas migration structures may represent a common phenomenon at convergent margins. This has implications in the overall understanding of the stress regime of the accretionary prism. Furthermore, the study of seafloor doming yields a better understanding of gas migration and the stability of the seabed. This is relevant to improve the safety of oil and gas exploration operations and provides an additional tool to understand the mechanics of upward migrating gas through marine sediments.

Zusammenfasssung

Fluidmigration und die damit verbundene markante Beschaffenheit der meeresbodennahen Sedimente ist ein häufiges und weitverbreitetes Phänomen an akkretionären Kontinentalrändern. Das Verständnis um Art, Variabilität und lateraler Verteilung von Fluidmigration ist von allgemeiner Bedeutung für den globalen Kohlenstoffkreislauf, für Naturgefahren (wie z.B. Meeresbodenstabilität) und der Kohlenwasserstoffexploration. Außerdem können wichtige Einblicke über die räumlichen sowie zeitlichen Prozesse des gesamten Subduktionssystems gewonnen werden.

In diesem Rahmen wurde eine kooperative Studie zwischen Neuseeland und Deutschland durchgeführt, um verschiedene, von vorherigen Arbeiten schon bekannte Gasaustrittsgebiete am Hikurangi Kontinentalrand vor Neuseeland, zu untersuchen. Während dieses NEMESYS-Projektes, wurden verschiedene geophysikalische, geochemische und mikrobiologische Messungen mit dem Forschungsschiff FS Sonne aufgenommen. Die hier vorgelegte Dissertation verwendet hauptsächlich geophysikalische Datensätze (2D, 3D seismische und hydroakustische Daten), mit zusätzlichen Informationen aus geochemischen Porenwasseranalysen.

Die verschiedenen Aspekte des Gasmigrationssystems von der Quelle bis zum Meeresboden werden durch drei Fallstudien näher beleuchtet. Die erste Studie kombiniert dafür 2D Mehrkanalseismik, seismische Geschwindigkeitsanalysen und geochemische Porenwasseranalysen, um den Gasmigrationsprozess zu untersuchen. Die wichtigsten Ergebnisse beinhalten die Erkenntnis, dass der Gasaustritt am Meeresboden fast ausschließlich durch Methan biogenen Ursprungs gespeist wird, ohne dass eine Wasserphase substantiell zum Transport (von z.B. gelöstem Methan) beiträgt, d.h. es existiert ein „trockenes Migrations-System". Die Gasquelle liegt dabei in einer maximalen Tiefe zwischen 1500-2100 m unterhalb des Meeresbodens. Das aufsteigende Gas wird hauptsächlich in vertikalen Aufstiegszonen fokussiert, und nur ein geringer Teil des Gases wird entlang von Sedimentschichten transportiert.

Die Verwendung des seismischen P-Cable Systems ermöglicht die 3D seismische Darstellung und damit eine räumliche Analyse der Fluidaustrittsstellen. Dadurch ist es zum ersten Mal gelungen, einen Zusammenhang der Fluidmigration mit dem tektonischen Spannungssystem im Akkretionskeil herzustellen. Die Studien ergaben, dass die Gasmigration durch ein vernetztes System von Bruchzonen erfolgt, wobei die Bruchzonen senkrecht zur Opouawe Bank Rückenachse orientiert

sind, und damit parallel zur Kompressionsrichtung der Subduktion. Diese, bisher nur selten beobachteten Bruchzonen sind das Ergebnis eines gravitativen Kollapses des gesamten Rückens. Die Bruchzonen entwickeln sich mit abnehmendem Auflastdruck in den oberen 100 m unterhalb des Meeresbodens zu eher zylindrischen Strukturen.

Der meeresbodennahe Teil des Gasmigrationssystems wurde in einer dritten Studie mit hydroakustischen und 2D seismischen Daten erforscht. Eine Abfolge verschiedener Entwicklungsstufen der Gasmigration wurde dabei offensichtlich. Diese Stufen beinhalten fokussierten Gasfluss, Gasakkumulation unterhalb relativ impermeabler Schichten, sowie Überdruckaufbau verbunden mit Sedimentaufwölbung. Die dafür benötigte Menge an Gas (in Form der Höhe der angestauten Gassäule) wurde anhand der Geometrie individueller Aufwölbungsstrukturen und unter Annahme repräsentativer, geomechanischer Sedimenteigenschaften berechnet. Diese Studie zeigt außerdem, dass sich Fluidaustrittsstellen (mit typischen chemo-synthetischen Meeresbodengemeinschaften) im Gegensatz zum vorherrschenden Verständnis, direkt aus Aufwölbungsstrukturen entwickeln können, ohne dass sich Meeresbodenvertiefungen (sogenannte Pockenmarken) am Meeresboden bilden.

Die Resultate der Studien lassen darauf schließen, dass die Beobachtung von durch Extension hervorgerufenen Gasmigrationspfaden durchaus ein viel weiter verbreitetes Phänomen an konvergenten Plattenrändern darstellen könnte, als bisher angenommen. Dies kann ebenfalls zu einer verbesserten Abschätzung der Spannungsverhältnisse im Akkretionskeil beitragen. Die Studie über Sedimentaufwölbung erweitert unsere Kenntnis der Zusammenhänge von Gasmigration und Meeresbodenstabilität. Relevant sind diese Beobachtungen u.a. für Erdöl- und Erdgasexploration, und sie bieten ein zusätzliches Instrument für das Verständnis der vertikalen Gasmigration durch marine Sedimente.

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List of abbreviations

2D	two-dimensional
3D	three-dimensional
AOM	anaerobic oxidation of methane
BGHS	base of the gas hydrate stability
BGHSZ	base of the gas hydrate stability zone
BSR	bottom simulating reflection
CDP	common depth point
CSEM	controlled source electro-magnetic
GC	gravity cores
GH	gas hydrate
GHSZ	gas hydrate stability zone
GI	generator-injector
IODP	Integrated Ocean Drilling Program
LVZ	low velocity zone
mbsf	meter below seafloor
MCS	multi-channel seismic
MUC	multiple corer
MWL	meteoric water line
PDB	Pee Dee Belemnite
sf	seafloor
SMOW	Standard Mean Oceanwater
TWT	two-way traveltime
VPDB	Vienna Pee Dee Belemnite
VSMOW	Vienna Standard Mean Oceanwater

1. Introduction

1.1. Motivations to study focused fluid flow

Fluid migration in marine sediments is a widespread and important geological process. The study of fluid flow systems has implications for understanding feedbacks between the global carbon reservoir, natural and exploration related hazards, seabed ecology, and global climate change. Methane (either of microbial or thermogenic origin) is the most abundant component of natural hydrocarbon gases within fossil fuel reservoirs (Claypool and Kvenvolden, 1983). The migration of fluids impacts the distribution of carbon in the subsurface and the amount seeping from the seabed (e.g. Berndt, 2005). At cold seeps and hydrothermal vents leaking hydrocarbons nurture a wide range of chemosynthetic biological communities (e.g. Judd and Hovland, 1988) which possibly represent the earliest microbial ecosystems on Earth (Martin et al., 2008). Hydrocarbons emitted at the seabed migrate through the water column and potentially reach the atmosphere, where the gases may contribute to a positive climate feedback. Methane, for example, has a 20-fold climate-warming potential than carbon dioxide (e.g. Shine et al, 1990). Major climate warming events, such as the late Permian-Triassic boundary ~253 Ma (Wignall, 2001; Retallack and Jahren, 2008), the Early Jurassic ~183 Ma (Hesselbo et al., 2000; Svensen et al., 2007), and the Paleocen-Eocene thermal maximum ~55 Ma (Dickens et al., 1995; Dickens, 1999, Zachos et al., 2001) were all associated with a sudden increase in the concentration of atmospheric methane.

At continental margins, the flow of fluids through sediments is a ubiquitous process. The interplay between fluid migration, gas hydrates and seafloor seepage has strong consequences for seafloor stability, but also impacts subduction zone processes such as megathrust fault locking and depth-distribution of earthquakes along the subducting plate (e.g. Ranero et al., 2008; Brasse et al., 2009; Worzewski et al., 2011). Active continental margins are very dynamic environments and marine geo-hazards are associated with natural events such as earthquake seismicity, slope failures, and gas venting. Trigger mechanisms for submarine slope failure are for example earthquake shaking, erosion or rapid sedimentation, but can also be connected to fluid transport and dissociation of gas hydrates where the liberation of fluids and free gas could yield substantial overpressure in the sub surface. The global gas hydrate reservoir represents a possible alternative energy resource as it contains a (though highly disputed) vast amount of methane gas (e.g. Kvenvolden and Lorenson, 2001; Milkov, 2004; Klauda and Sandler, 2005; Wallmann et al., 2012; Piñero et al., 2013). Subduction margins contain a significant portion of that estimated global gas hydrate inventory (e.g. Kastner, 2001). The mechanics of fluid migration at convergent margins is key to understand the spatial and temporal variation of geological processes associated with the subduction system (e.g. Hensen et al., 2004; Ranero et al., 2008). Fluid migration systems can provide insights into processes of tectonic deformation, compaction of the sedimentary sequences, and the reduction of porosity. The study of cold seep may provide insights into deep-seated processes (Kastner et al., 1991; Hensen et al., 2004).

1.2. The SO-214 NEMESYS Project

The work presented in this thesis was carried out in the framework of the NEMESYS (**Ne**w Zealand **me**thane **syst**em) project, funded by the German Federal Ministry for Education and Research (BMBF) under grant 03G0214A. In cooperation between the Leibniz Institute of Marine Science, IFM-GEOMAR (now GEOMAR, Helmholtz Centre for Ocean Research Kiel) and the Federal Institute of Geosciences and Natural Resources (BGR) and with partners from the Institute of Geological and Nuclear Sciences (GNS Science) as well as the Royal Netherlands Institute for Research (NIOZ), the cruise SO214, with the German research vessel Sonne took place from 9th of March until the 22nd of April, 2011.

The NEMESYS project aimed to extend the understanding of cold vents and their variability on the Hikurangi Margin off the east coast of New Zealand's North Island. The vent sites occur within the region of gas hydrate stability; thus, understanding the dynamics of fluid migration and hydrate formation are crucial to assess seafloor stability. Based on the results of previous expeditions (Greinert et al., 2010a; Bialas, 2011) and the preceding New Vents project (Bialas, 2007), five working areas (Fig. 1.1) were selected for expedition SO214. The objectives were to improve the knowledge of seep sites at active continental margins and review present models of the buildup of seeps and their feeder structures with respect to the tectonic regime.

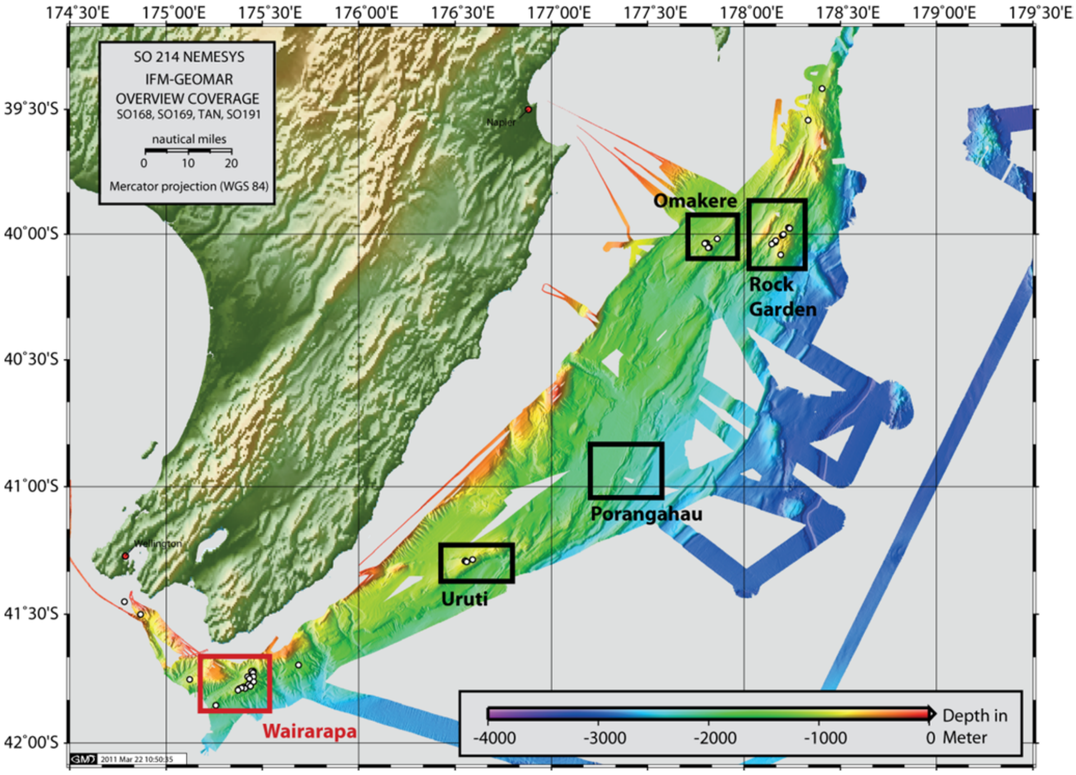

Figure 1.1: Overview map of the Hikurangi Margin, offshore New Zealand's North Island, displaying the working areas (rectangles) of the cruise SO214 Nemesys. Wairarapa area, the focus of this thesis study, is marked with a red rectangle; white dots mark known seep sites (modified after Bialas 2011).

The project pursued a multi-scale approach with geophysical methods and geological, geo-chemical and biological analysis to study fluid sources, migration pathways, seep habitats and the impact on the water column. The focus of this thesis is Opouawe Bank in the Wairarapa study area with the main emphasis on the geophysical survey. Opouawe Bank was selected for its dense occurrence of cold seeps with feeder channels of various expressions as seen in existing 2D seismic data. The spatial structure of the observed fluid migration pathways could not be determined from the existing 2D data. Consequently, a 3D seismic survey was acquired to image lateral dimensions with the purpose to evaluate the formation processes of the fluid migration systems.

1.3. Data

The geophysical data of the NEMESYS project presented in this thesis include 3D and new 2D multi-channel seismic (MCS) data and coincidently recorded high-resolution Parasound data. The acquisition of the 3D seismic volume was constrained by expected profiling time, the optimal coverage of seep sites and the required minimum inline distance for the 3D volume. As a result, the 3 by 8 km wide area covered with seismic data comprises five seep sites (Piwakawaka, Riroriro, Pukeko, North Tower, and South Tower). The 2D seismic lines cover 3 additional seep structures (Takahe, Takapu, and Papango).

The 3D seismic data were recorded with the P-Cable system (Fig. 1.2; Planke and Berndt, 2003) from GEOMAR, Kiel, which provides 3D data and has a much higher resolution than conventional 3D data used in hydrocarbon exploration surveys. The acquisition system consists of parallel streamer segments mounted to a cross cable. The cross cable is towed perpendicular to the heading of the vessel and spread by two trawl doors. Attached to the trawl doors are GPS antennas for navigation (Bialas, 2011). During cruise SO214, the P-Cable system consisted of 16 parallel-towed digital streamers, each with 8 groups of 2 hydrophones and a group distance of 1.5 m.

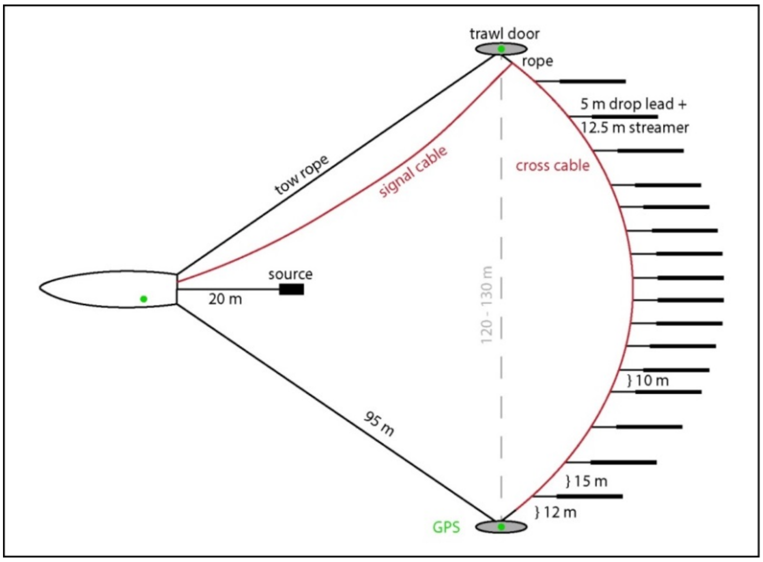

Figure 1.2: Sketch of the P-Cable design applied during SO214 with 16 parallel towed streamer-segments. A GI-gun was used as source. GPS was recorded on the trawl doors and on a reference antenna mounted to the ship (green dots).

For 2D MCS profiling, the 16 individual streamer segments used for the 3D survey were assembled into a continuous 200 m long digital streamer, with 128 channels at the same group spacing as above. Three birds for depth control were mounted to the assembled 2D streamer. A single generator-injector (GI) gun from Sercel Marine Sources Division, with a volume of 3.8 l (105 in^3 generator chamber, 105 in^3 injector chamber) was the acoustic source for both 2D and 3D seismic data. The gun was operated at a shot interval of 6 s at an average ship speed of 3 kn. The MCS and the P-Cable datasets both have a usable bandwidth (after frequency filtering) of 50 – 300 Hz.

The main processing steps of 3D seismic data included navigation correction for the 16 streamers and the source, trace editing, frequency filtering, and binning to a 3D grid with a nominal cell size of 3.125×3.125 m. A water velocity stack was the input to 3D Kirchhoff time migration, due to insensitivity of moveout velocities from short offsets. The 2D single-streamer data processing included navigation processing, trace editing, frequency filtering, velocity filtering, and a crooked line binning with cell size of 1.5 m. After water velocity stack, a post-stack Kirchhoff time migration with a representative velocity function below the seabed was applied.

The Parasound is an echosounder system using the parametric effect, a nonlinear acoustic interaction of two similar frequencies (e.g. 18 kHz, 22 kHz), which produce a signal with the difference-frequency (e.g. of 4 kHz). The advantage of the system is a controlled repeatable wavelet and a small emission cone, as the generated signal travels within the 4° angle of the original high frequency waves. The system is permanently installed on the vessel and was used to map gas flares in the water column (with the primary high frequency at 18 kHz) and to image the sediment column (with the secondary low frequency at 4 kHz). The depth penetration was up to 100 m below the seafloor with a high resolution at decimeter scale.

1.4. Fluid flow in marine sediments

Fluid flow refers to migration of liquids and gases and is a widespread natural process in marine sediments that has considerable implications for the marine environment and impacts the sediment, the seabed, and the water column. Documented for a variety of oceanographic settings, fluid flow occurs from the deep ocean to coastal areas and for different tectonic settings encompassing convergent, divergent, and transform plate boundaries, as well as intraplate setting (e.g. Lewis and Marshall, 1996; Whiticar, 2002; Dupré et al, 2010; Gay et al., 2012; Talukder, 2012; Leduc et al., 2013)

Depending on the processes inducing fluid production and migration, diverse kinds of fluid flow systems exist including hydrothermal vents (Jamtveit et al., 2004; Planke et al., 2005), cold seep systems (Sibuet and Olu-Le Roy, 2003; Talukder, 2012), polygonal faults (Cartwright, 1994; Berndt et al., 2003) and mud volcanos (Dimitrov 2002; Kopf 2002). The types of fluid flow are affected by the fluid composition and the tectonic setting. Berndt (2005) differentiates between volcanism-controlled, compaction-controlled, fresh-water and petroleum systems. Though diverse fluid flow systems with different seabed expressions exist, they all exhibit analogies in their general structure, e.g. with a fluid reservoir or source, a plumbing (feeder) system transporting the fluids, and corresponding seabed expressions (e.g. Judd and Hovland, 2007; Talukder, 2012).

Controls on focused fluid flow involve the geological environment at the source-region of the fluids, chemical fluid composition, triggers and driving forces, and the possible existence of preferential migration pathways. In porous media, rates of fluid flow are controlled by Darcy's law, which, in principal, describes that the permeability and pore pressure difference of the fluid flow system affect the amount of fluids passing through. Systems of low permeability may induce fractures and faults by generation of overpressure, where rates of fluid flow are dependent on the effective permeability of the conduits (e.g. Fisher et al., 2003; Jain and Juanes, 2009; Judd and Hovland, 2007 and references therein). Advection and diffusion are the common mechanisms for vertical and lateral fluid migration and are essential to explain for example the transport of methane into the gas hydrate stability zone (Liu and Flemings, 2007).

The driving forces for fluid migration in continental margins are excess pore fluid pressure (overpressure) and buoyancy (e.g. Anketell et al., 1970; Hart et al., 1995; Osborne and Swarbrick, 1997). Buoyant fluids rise towards the surface; this includes flow through permeable formations, along discontinuities and the vertical migration of gas by propagating fractures and voids in gas chimneys. Buoyancy driven fluid migration will cease if a neutral level of buoyancy is reached or a barrier prevents further fluid ascent. Permeability barriers may result in overpressure accumulation and represent a seal (Cartwright et al., 2007). Excess pore pressure can also result from the reduction of the pore volume, through compaction or compression and volume changes (Osborne and Swarbrick, 1997), for example faster gas production than dissipation. Overpressure will accumulate in the pore space underneath a seal until it exceeds the resistance to capillary entry pressure or fracture failure (Clayton and Hay, 1994). Related to excess pore pressure fracturing, plastic deformation, fluidization, sediment doming and explosive decompression may occur.

1.4.1. Cold seep systems

Cold seeps are sites where (in contrast to hydrothermal systems) cold fluids are transported from the sediment to the ocean and thereby sustaining chemosynthetic ecosystems at and near the seabed. The migration of fluids is generally linked to tectonic deformation, compaction, porosity reduction of the sedimentary sequence and mineral dehydration. Cold seeps occur at active and passive continental margins and also transform plate boundaries (Fig. 1.3, e.g. Hovland and Judd, 1988; Suess, 2014; and references therein). Active continental margins are either accretionary or erosional. Accretionary subduction zones form by sediments off-scraping from the subducting oceanic crust and produce an accretionary wedge with ridges and sedimentary basins, which are the main source of the seep fluids. In contrast, at erosional subduction zones a large portion of the sediment cover gets subducted together with the oceanic plate and thus undergoes a different diagenetic pathway than the accreted sediments (von Huene and Scholl 1991). At passive margins fluid flow is mainly driven by the sediment load, differential compaction, facies changes and overpressure (e.g. Berndt, 2005; Suess, 2010). Cold seep systems are windows into processes of the sub-seafloor and provide for example clues of the source depth, fluid and sediment interaction during ascent, and fluids from clay-dehydration provide criteria for source temperature and depth (Kastner et al., 1991; Martin et al., 1996; Hensen et al., 2004; Suess, 2014; Riedel et al., 2010).

Cold seep systems comprise a plumbing system, where fluids migrate from the source to the seabed, and seepage features at the seafloor or in the shallow sediment (Talukder, 2012). The plumbing system is associated with structural (along fractures and faults) and stratigraphical (e.g. along bedding planes) fluid migration pathways. The internal structure and driving mechanisms of these plumbing systems are the least understood components of fluid flow systems (e.g., Judd and Hovland, 2007; Anka et al., 2012). Cold seeps are often connected with vertical focused fluid flow conduits that crosscut the sedimentary strata, which are typically described as chimneys or pipes, based on their appearance on seismic data (Riedel et al., 2002; Cartwright et al., 2007; Løseth et al., 2009; Andresen, 2012). Chimneys refer to seismically dimmed or wipe-out zones and may reach diameters of several km (e.g. Tommeliten, Løseth et al., 2009) and pipes are typical narrow vertical seismic anomalies. Recent overview publications characterize these fluid conduits as seal bypass systems (Cartwrigth et al., 2007), hydrocarbon leakage systems (Løseth et al., 2009) and hydrocarbon plumbing systems (Talukder, 2012; Andresen, 2012).

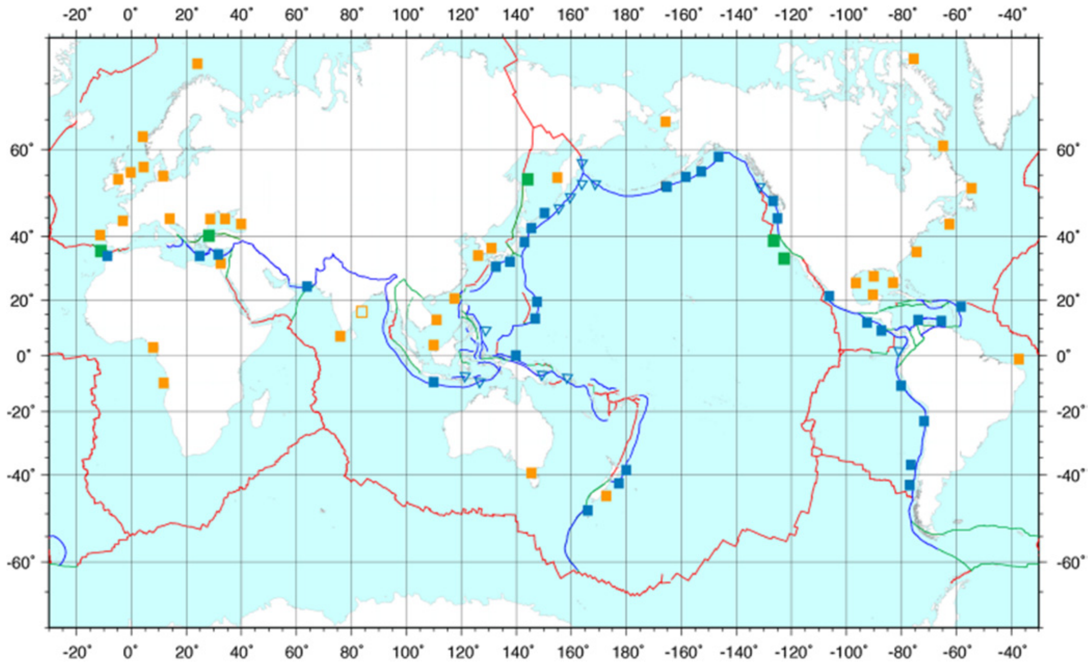

Figure 1.3: Map of global cold seep occurrence (from Suess, 2014) with sites at passive margins incl. groundwater seeps (orange), active margins (blue), and transform margins (green).

Seafloor expressions related to cold seep activity are various: Seeping fluids may produce mounds, pingo-like features domes, mud volcanoes, or negative morphological expressions, known as pockmarks (Judd and Hovland, 2007). Mud volcanoes are elevated morphologic features generated from ejected and extruded sediments (Kopf, 2002). Seabed domes of a small vertical extent are inferred to correlate with overpressure build-up of accumulating gas (Hovland and Judd, 1988; Barry et al, 2012) and potentially lead to the development of pockmarks (Hovland and Judd, 1988). Pingo-like features are believed to result from shallow gas hydrate formation (Hovland and Svensen, 2006; Paull et al., 2008).

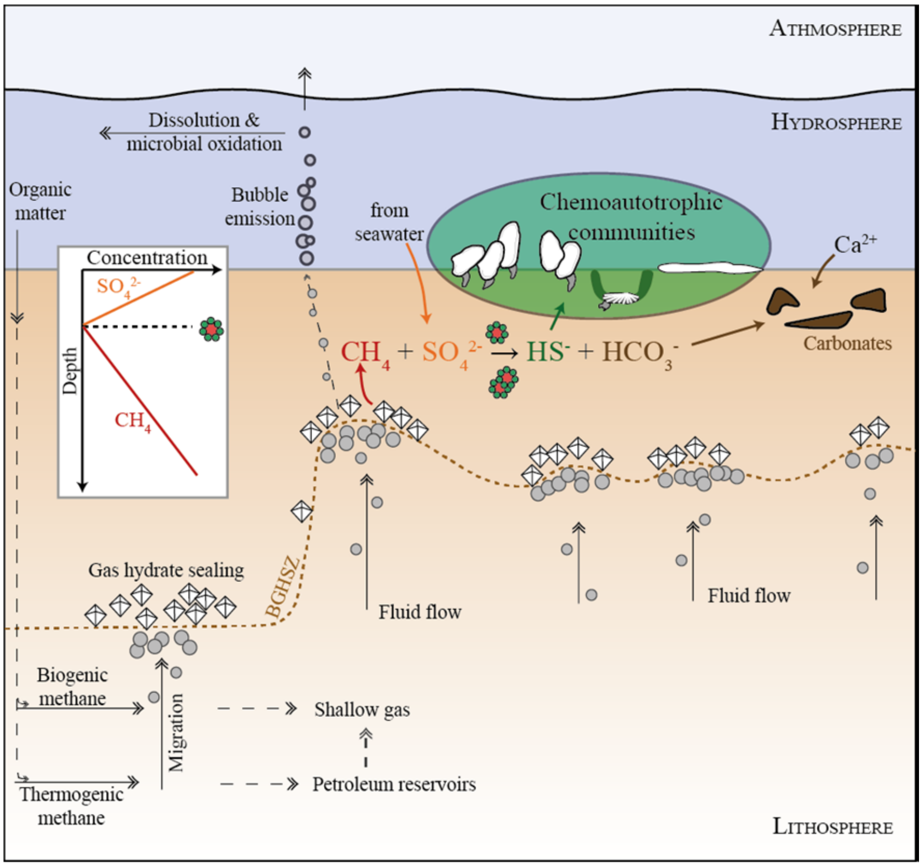

Figure 1.4: Schematic overview of processes in the sediments at cold seep systems (from Roemer, 2011). The diagram shows the principle cycle from the sources to the fate of methane, and the processes leading to the consumption of dissolved methane by AOM, which fuels the chemoautotrophic communities and induces carbonate precipitations.

The dominant microbial process at cold seeps is the anaerobic oxidation of methane (AOM) coupled with sulfate reduction (Fig.g 1.4). Boetius et al. (2000) identified a consortium of methanotrophic archaea to mediate AOM. The subsequent formation of hydrogen sulfide serves as energy source for chemosynthetic organisms colonizing the seabed, such as bacterial mats, clams, mussels, and tubeworms (e.g. Sahling et al., 2002). Bacterial mats were initially discovered offshore California in the Santa Barbara Channel (Spies and Davies, 1979). AOM is an important sink for methane and limits the amount transferred to the water column and thus impacts on the global carbon cycle. The precipitation of carbonates is also a result of microbial activity at seep sites. Authigenic carbonates are potential recorders of fluid composition and venting activity (e.g. Teichert et al., 2003; Judd and Hovland, 2007) and they form small chimneys, crusts and concretions.

1.4.2. Methane gas in marine sediments

The term "fluid" encompasses liquids and gases and commonly comprises pore water and hydrocarbons. Primarily associated with fluid flow systems and the formation of gas hydrates is methane gas. It can originate from biogenic (referring to microbial and thermogenic gas) and abiogenic processes (e.g. Schoell, 1988; Welhan, 1988; Whiticar, 1999).

Biogenic methane is generated through the consumption of organic matter by microbes (e.g. Schoell, 1988). High amounts of organic matter are often related to areas of high sediment input. Microbial methane is produced and consumed under anoxic conditions, by CO_2 reduction or fermentation (Schoell, 1980; Whiticar et al., 1986). Thermogenic methane results from the transformation of organic matter under high temperatures above 80°C through thermochemical reactions. Abiogenic methane is derived from inorganic processes by thermochemical reactions (e.g. Schoell, 1988) and occurs at hydrothermal systems (Welhan, 1988).

The origin of the methane can be characterized with stable isotope analysis of carbon (C) and hydrogen (H). The resulting isotope signatures (reported in the standard δ-notation, e.g. $\delta^{13}C$; $\delta^{2}H$ or δD) from the partitioning of light and heavy isotopes of C (^{12}C and ^{13}C) and H (^{2}H and ^{1}H) give insight into the methane generation (e.g. Rice and Claypool, 1981; Schoell, 1988; Whiticar, 1999). Microbial gas possesses a lighter methane isotopic value (Fig. 1.5), compared to the heavier carbon isotope values typical for thermogenic methane (e.g. Whiticar, 1999). Furthermore, the proportion of methane in a hydrocarbon gas mixture differs for thermogenic and microbial gas. During microbial processes methane is produced at a significantly greater rate than higher molecular weight hydrocarbons such as ethane, propane, and butane, thus the formation processes can be deduced by combining the information of molecular weight and isotopic composition (Fig. 1.6, Bernard et al., 1976; Whiticar, 1999).

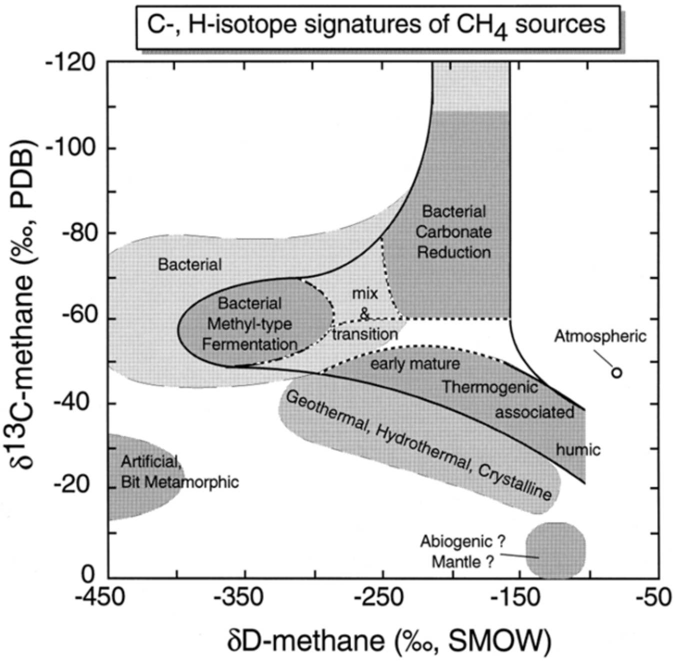

Figure 1.5: Cross plot of δ^{13}C and δD values of methane (from Whiticar et al., 1999) for the classification of thermogenic and biogenic generated methane (PDB: Pee Dee Belemnite, standard for ^{13}C; SMOW: Standard Mean Oceanwater).

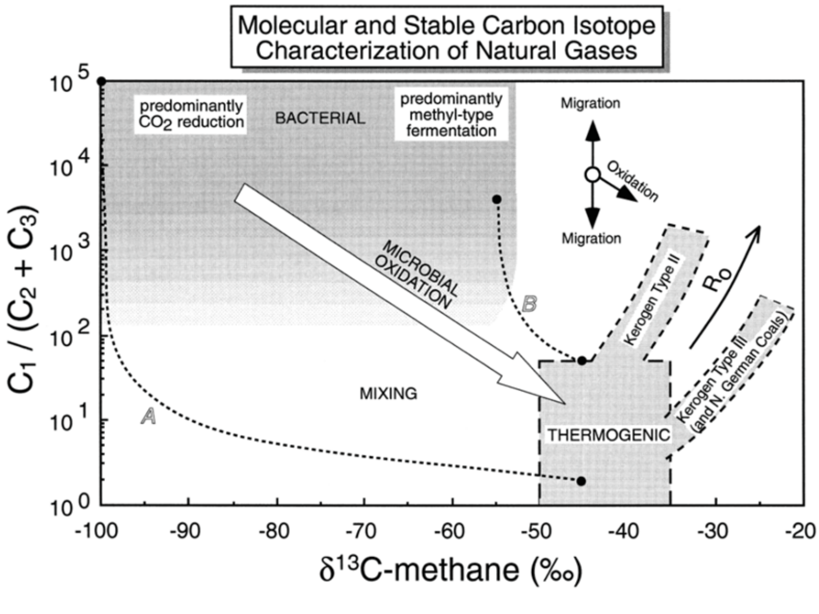

Figure 1.6: Bernard diagram (from Whiticar, 1999) combining the molecular (C1/(C2+C3)) and isotopic (δ13C) compositional information and displaying biogenic or thermogenic origin of natural gases or mixing of both. The arrows indicate the relative compositional effects of oxidation or migration.

1.4.3. Gas hydrate

Gas hydrates are clathrates, a cage structure made up of hydrogen-bonded water molecules and an enclosed guest molecule (e.g. Kvenvolden, 1993; Sloan and Koh, 2008). The guest molecule is a low molecular weight gas, typically methane. Three types of natural gas hydrate are known (structures I, II, and H) and the amount of methane that is incorporated depends on the structure (e.g. Sloan, 2003). The most common is structure I, where CH_4 is the main hydrate-forming gas (Fig. 1.7), with a high capacity to store gas, as 1 m^3 of hydrate contains as much as 164 m^3 of methane at standard temperature and pressure conditions (Kvenvolden, 1993). Larger gas molecules, such as ethane or propane can be included in the larger cages of structure II and H (Kvenvolden, 1995; Sloan, 2003).

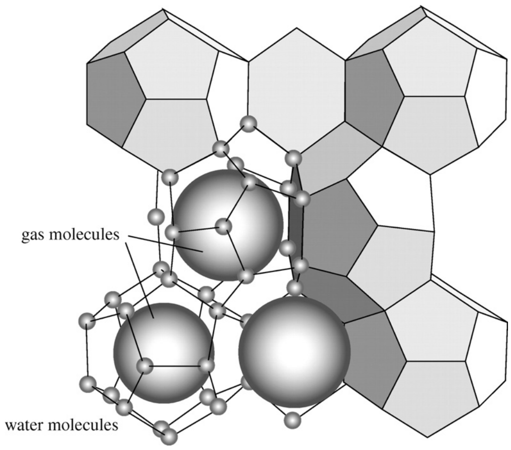

Figure 1.7: Structure I gas hydrate, the linked water molecules form a cage trapping a gas molecule such as methane within (Maslin et al., 2010 and references therein).

Gas hydrate stability depends (among other factors) on temperature and pressure (Fig. 1.8). Thus gas hydrate occurrence is restricted to a few global environments: The polar regions, where it is associated with permafrost, deep marine environments, such as continental margins (Beaudoin et al., 2014), and sedimentary units beneath deep lakes (e.g. Khlystov et al., 2013). Along continental margins gas hydrates usually occur in the upper few hundred meters below the seafloor at more than 300 – 500 m water depth (e.g. Kvenvolden, 1993; Ruppel, 2007; Sarkar et al., 2012)

The extent of gas hydrate stability is controlled by the relationship between the gas hydrate phase boundary and the temperature-pressure regime of a specific setting. In marine settings local geology, sea bottom temperature, gas composition, and pore water salinity can also affect the stability of gas hydrate (Kvenvolden, 1993). The intersection of the phase boundary and the geothermal gradient defines the base of the gas hydrate stability (BGHS) and the regional gas hydrate stability zone (GHSZ) is the section between the BGHS and the seafloor. Below the GHSZ free gas may exist. The impedance contrast between gas hydrate bearing sediments (relatively high P-wave velocity) and free gas (low P-wave velocity) produces the bottom simulating reflection (BSR) in seismic reflection data (first identified by Shipley et al., 1979, on data around the Blake Ridge, offshore Carolina, US). Therefore gas hydrate provinces in marine sediments can directly be inferred from BSRs.

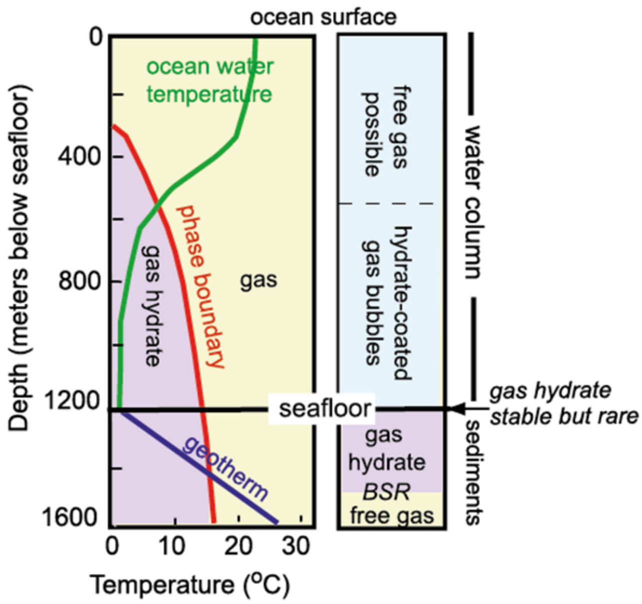

Figure 1.8: Pressure-temperature conditions and methane hydrate stability in marine settings (from Ruppel, 2007).

1.5. Fluid flow studies at the Hikurangi margin

At the Hikurangi Margin plenty indications for fluid flow exist, including oil, gas, and mud expulsion sites on land (Field et al., 1997), and cold vent sites offshore, which were initially inferred from fishery by-catch in 1989 (Lewis, 1990). Exhumed onshore seep-carbonates show that seafloor seepage is ongoing since at least 22 Ma years (Campbell et al., 2008). Over the last two decades many methane seep sites have been discovered on the margin (e.g. Lewis and Marshall, 1996; Greinert et al., 2010a). They are associated with chemosynthetic fauna (e.g. Lewis and Marshall, 1996, Faure et al., 2010; Baco et al., 2010; Bowden et al., 2013), carbonate precipitation (e.g. Liebetrau et al., 2010), acoustic anomalies in the water column (e.g. Greinert et al., 2010b), fluid migration pathways (chimneys and pipes) within the sediment (e.g. Netzeband et al., 2010, Crutchley et al., 2010a), and a wide-spread BSR marking the BGHS (e.g. Henrys et al., 2003).

Most of the seep sites are located at the crests of accretionary ridges in the central part of the Hikurangi margin. At the southernmost expression of the Tonga-Kermadec-Hikurangi subduction zone, the 25 Myr old active Hikurangi margin east of New Zealand's North Island has formed in response to the westward subduction of the Pacific Plate beneath the Australian Plate. The subduction rate is 49 mm/yr at 37°S, declines southwards to 40 mm/yr at 42°S and southwest of 42°S strike slip motion begins to dominate (DeMets et al., 2010; Collot et al., 1996; Beavan et al., 2002; Barnes et al., 2010). While the northern margin (off Raukumara Peninsula) is dominated by tectonic erosion (Lewis et al., 1998; Collot et al., 2001), the central margin off the Wairarapa coast is a classical imbricated thrust wedge characterized by accretion (Lewis and Pettinga, 1993; Collot et al., 1996; Barnes and de Lépinay, 1997). Per year a volume of 24 m^3 of fluid per meter strike length (Townend, 1997) is added by frontal accretion to the Hikurangi margin. Consequently the abundant evidence of gas escaping offshore and onshore is released by subsequent compaction and tectonic deformation.

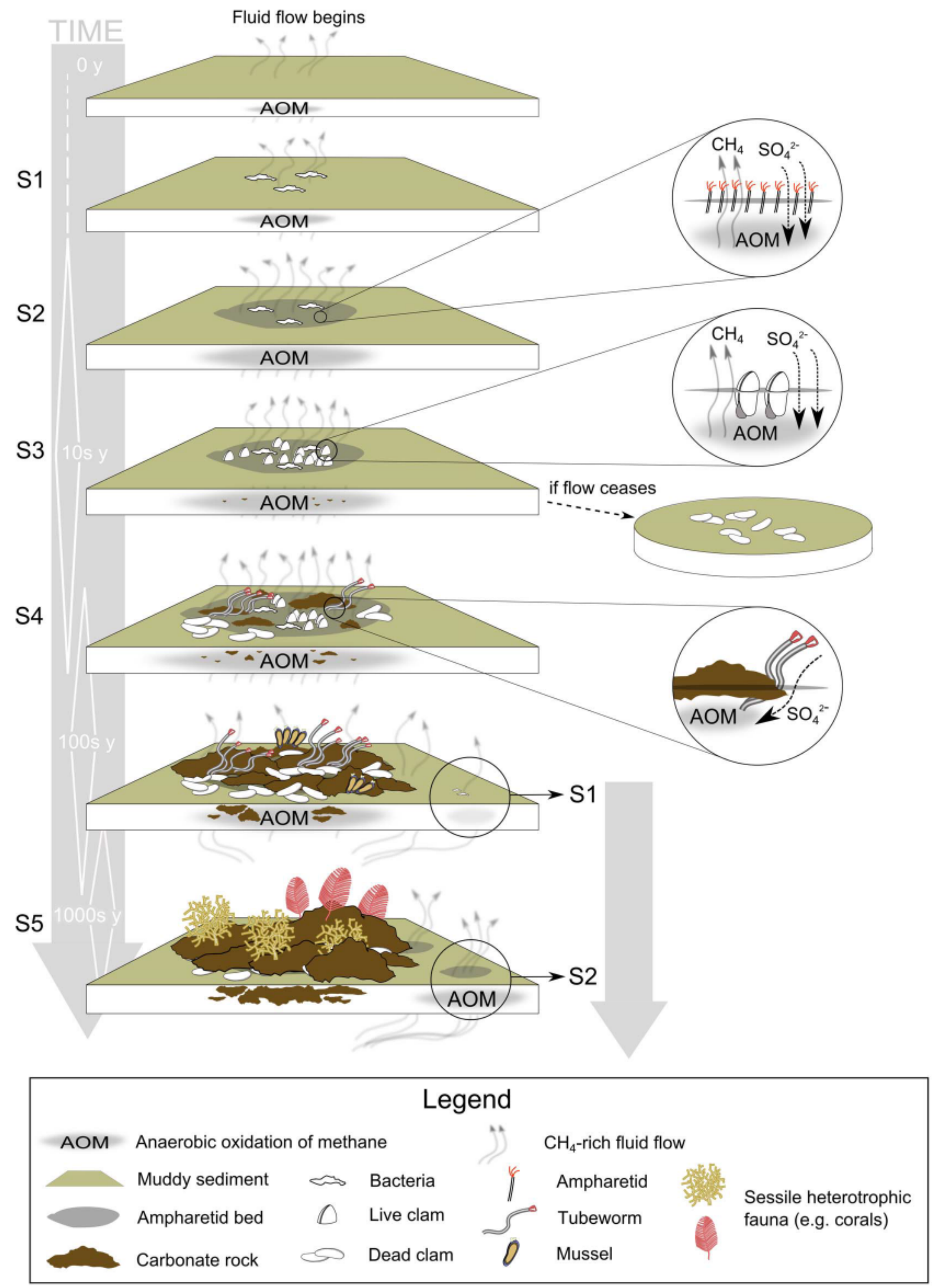

Figure 1.9: Hypothetical succession sequence of cold seep site communities at the Hikurangi Margin (from Bowden et al., 2013).

Related to the seep sites is a widespread gas hydrate province along the margin. It extends from -38.5° S to -42.5° S along the margin and from water depths of 600 m eastwards, is inferred from BSRs on seismic data (e.g. Henrys et al., 2003; Henrys et al., 2009; Greinert et al., 2010b; Bowden et al., 2013). Diffusive upward migration of methane presumable causes a semi-continuous BSR (Kroeger et al., 2015). BSR strength has been used to illustrate the distribution and potentially highly concentrated deposits of gas hydrates (Henrys et al., 2009). Heat flow on the margin estimated from BSR depths reveal low values of 44 ± 10 mW m^{-2} (e.g. Townend, 1997; Henrys et al., 2003; Pecher et al., 2010), with maximum values at the toe of the accretionary prism and decreasing landwards from the deformation front. At the flanks of the ridges high heat flow values combined with low heat flow values at the anticlines illustrate the topographic effect on subsurface temperatures (e.g. Henrys et al., 2003) and in the vicinity of seep site elevated heat flow values occur (Krabbenhoeft et al., 2013). The interaction of gas hydrate stability with tectonic uplift of a ridge (Rock Garden) was proposed to cause sediment weakening and leading to seafloor erosion (Pecher et al., 2008; Ellis et al., 2010). Shallow deposits of gas hydrate, close to the seafloor at Rock Garden, may produce a permeability reduction, partially trapping fluids, which leads to overpressure accumulation and thus a weakening of near seafloor sediments (Crutchley et al., 2010b).

Extensive evidence for the migration of free gas into the GHSZ is reported for the Hikurangi margin, occurring along sedimentary layering across the phase-boundary of the BSR (e.g., Crutchley et al., 2010a; Crutchley et al., 2015) and across strata through subvertical chimney or pipe structures (e.g., Netzeband et al., 2010; Krabbenhoeft et al., 2013; Plaza-Faverola et al., 2014). The migration pathways through the GHSZ to the seabed are structurally or stratigraphically controlled (Krabbenhoeft et al., 2013).

Methane emissions at the seabed have been shown to be temporally variable and correlate with tides and currents (Krabbenhoeft et al., 2010; Klaucke et al., 2010). The seep sites along the margin feature different seabed morphologies with carbonate crusts or blocks, carbonate free seabed, and different faunal communities indicating successive stages of seep development (Fig. 1.9; Bowden et al., 2013; Dumke et al., 2014). Geochemical analyses of authigenic carbonate samples recovered from three seep areas (Opouawe Bank, Uruti and Omakere Ridge, compare Fig. 1.1) show different phases of seep activity between 12,400± 160 and 2090± 850 years BP and intermediate intervals between 4950± 650 and 3960± 50 years BP (Liebetrau et al., 2010).

1.6. Thesis Structure

This thesis consists of an introductory chapter (Chapter 1) and three case studies that describe the diverse parts of the gas migration system at Opouawe Bank, offshore New Zealand (Chapters 2-4), and a concluding chapter (Chapter 5). The case studies represent stand-alone manuscripts with their own abstract, introduction, methods, results, discussion, and conclusion sections. They will be submitted (Chapter 3) and have been published (Chapter 2 and 4) in international peer-reviewed journals.

Chapter 2 demonstrates the strength of linking geochemical and geophysical data to investigate the nature of gas migration beneath the seafloor. We show that the cold vents at Opouawe Bank, offshore New Zealand are fueled by the seepage of biogenic methane gas and provide an insight on how gas migration may affect the gas hydrate system. Accumulating free gas underneath the BGHS is generated at a maximum depth of 1500-2100 m below the seafloor. Gas migration through Opouawe Bank is influenced by anticlinal focusing and several focusing levels within the BGHS and takes place along structurally controlled conduits and along dipping strata across the BGHS.

This chapter is published by the journal Geo-marine Letters as *Koch, S., Schroeder, H., Haeckel, M., Berndt, C., Bialas, J., Papenberg, C., Klaeschen, D., and Plaza-Faverola, A. (2016) Gas migration through Opouawe Bank at the Hikurangi Margin, offshore New Zealand. Geo-marine Letters 36, 187-196*. Supporting Information can be found in the Appendix A.1.

Contributions to Chapter 2: I carried out seismic analyses, prepared illustrations, and wrote the first draft of the manuscript. Matthias Haeckel contributed with the analysis and discussion of the geochemical data. Henning Schroeder and Dirk Klaeschen provided the velocity analysis based on the reprocessing of the line Pegasus_19, which was originally processed by Andreia Plaza-Faverola and Dirk Klaeschen. C. Papenberg and D. Klaeschen processed the seismic data of the NEMESYS project. All co-authors and external reviewers helped improving and revising the manuscript.

Chapter 3 deals with the local stress conditions within Opouawe Bank, on the Hikurangi subduction zone and the implications for the gas migration pathways. We show that gas migrates along extensional fractures through the accretionary ridge. The ridge is build up by compressive stress and secondary longitudinal extension. This enables the development of elongated pathways parallel to the most compressive stress and a transition in the upper 100 m of the sediment, to pipe-like structures due to declining differential stress. The observed gas migration structures are singular, but as only limited high-resolution 3D data is available, they might be more common than thought for subduction zones and accretionary ridges.

This chapter will be submitted under the title *Elongated fluid flow structures: Stress control on gas migration in Opouawe Bank, New Zealand* by S. Koch, C. Berndt, G. Crutchley, D. Klaeschen, and J. Bialas to a peer-reviewed journal.

Contributions to Chapter 3: I analyzed the data and wrote the manuscript, with the contribution of C. Berndt, G.J. Crutchley, and J. Bialas through intensive discussions. D. Klaeschen, G.J. Crutchley, J. Bialas, and S. Koch contributed to the acquisition of the seismic and Parasound data. J. Bialas led the survey. D. Klaeschen processed the seismic data. All co-authors helped improving and revising the manuscript.

Chapter 4 presents a study on seafloor doming and combines both theory and field data to provide evidence for the validity of the theory of seabed doming and associated processes. Based on geo-mechanical quantifications, gas column heights necessary to create different dome geometries were calculated with respect to examples from Opouawe Bank, New Zealand. The process of sediment doming is considered in the analysis of evolutionary stages of seep formation. Our results suggest that the progression from channeled gas flow to gas trapping results in overpressure build-up in the shallow sediment, which leads to doming and finally by breaching of domed seafloor sediments a new seep site can develop.

This chapter is published in the journal Geology as *Koch, S., Berndt, C., Bialas, J., Haeckel, M., Crutchley, G.J., Papenberg, C., Klaeschen, D., and Greinert, J. (2015). Gas-controlled seafloor doming. Geology 43, 571-574*.

Contributions to Chapter 4: S. Koch, G.J. Crutchley, C. Papenberg, D. Klaeschen, and J. Greinert contributed to the acquisition of the seismic and Parasound data. J. Bialas led the survey. G.J. Crutchley and J. Greinert processed the sediment echosounder data. C. Papenberg and D. Klaeschen processed the seismic data. I analyzed the data and wrote the manuscript, with the contribution of C. Berndt, J. Bialas, M. Haeckel and G.J. Crutchley through intensive discussions. All co-authors, and four external reviewers helped improving and revising the manuscript.

Chapter 5 summarizes the findings of the three previous chapters and draws integrated conclusions for the overall motivation of this study. Further, I provide an outlook on future work and follow-up studies.

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2. Gas migration through Opouawe Bank

At the Hikurangi margin offshore New Zealand

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2.1. Abstract

This study presents 2D seismic reflection data, seismic velocity analysis, as well as geochemical and isotopic porewater compositions from Opouawe Bank on New Zealand's Hikurangi subduction margin, providing evidence for essentially pure methane gas seepage. The combination of geochemical information and seismic reflection images is an effective way to investigate the nature of gas migration beneath the seafloor, and to distinguish between water advection and gas ascent. The maximum source depth of the methane that migrates to the seep sites on Opouawe Bank is 1,500 – 2,100 m below seafloor, generated by low-temperature degradation of organic matter via microbial CO_2 reduction. Seismic velocity analysis enabled identifying a zone of gas accumulation underneath the base of gas hydrate stability (BGHS) below the bank. Besides structurally controlled gas migration along conduits, gas migration also takes place along dipping strata across the BGHS. Gas migration on Opouawe Bank is influenced by anticlinal focusing and by several focusing levels within the gas hydrate stability zone.

2.2. Introduction

Cold seeps and associated gas migration through marine sediments occur worldwide on continental margins (e.g., Hovland and Judd, 1988; Sibuet and Olu, 1998; Cartwright et al., 2007), and are often linked to gas hydrate deposits. Investigation of the processes related to gas migration and accumulation in marine sediments is driven by different motivations, such as the general understanding of chemosynthetic metabolism, its role in the subduction zone factory, and also economic interest in the exploitation of gas and gas hydrate resources. The global gas hydrate reservoir contains a huge amount of methane, and thereby represents an alternative energy resource (e.g., Kvenvolden, 1988; Milkov and Sassen, 2002; Wallmann et al., 2012; Pinero et al., 2013).

Furthermore, determining the fluid flow system of a convergent continental margin is a key to understanding the spatial and temporal variations of geological processes associated with the subduction system (e.g., Hensen et al., 2004; Klaucke et al., 2008; Ranero et al., 2008). For this, it is important to identify the origin of gas to decide if seeps can be used as a window into deep-seated processes, and whether subduction zone mechanisms can be derived from the study of seep sites (Hensen et al., 2004). Combining geochemical information and seismic reflection images is an effective way to investigate the nature of gas migration beneath the seafloor (e.g., Brooks et al., 2000).

Gas migration is often accompanied by anomalous high-amplitude reflections, as well as by acoustic blanking or turbidity of the seismic signal. Columnar zones of seismic blanking and turbidity, referred to as chimneys or pipes, are caused by absorption and scattering of acoustic energy by gas-charged sediments (Gay et al., 2007; Løseth et al., 2009; Cartwright and Santamarina, 2015). Highly focused and energetic fluid flow can lead to suppressed reflectivity, as migrating fluids physically destroy the stratified fabric of the sediments through which they migrate (Wood and Gettrust, 2001; Gorman et al., 2002). High-amplitude anomalies can occur, for example, when gas is present in the sediments, reducing the seismic velocity and increasing the impedance contrast to the adjacent sediments (Løseth et al., 2009), or they may result from a strong acoustic impedance contrast by hydrocarbon-related diagenetic zones (O'Brien and Woods, 1995).

Fluid flow toward seep sites in marine sediments often has to penetrate and migrate through the gas hydrate stability zone (GHSZ). The specific seismic anomalies at the base and within the GHSZ provide useful information on the processes involved in transporting fluids through the hydrate stability zone. The most widespread and easily recognized gas hydrate-related reflection in marine sediments is the bottom simulating reflection (BSR), a high-amplitude reflection with a negative impedance contrast. The BSR marks the base of the GHSZ, as partially gas hydrate saturated sediments overlie sediments containing free gas (Holbrook et al., 1996). Variations in the appearance of the BSR can indicate disturbances in the hydrate system for example, gas chimneys penetrating the base of gas hydrate stability (BGHS), shoaling of the BSR due to locally increased heat flow, or gas migration along sedimentary layering that transects the BSR (e.g., Pecher et al., 2010; Crutchley et al., 2014).

Cold vent ecosystems at the sediment surface are fueled primarily by methane generated from buried organic matter either thermo-catalytically at temperatures above 80 °C or microbially at lower temperatures. This methane is transported to the seafloor by upward migration of the buoyant gas phase or dissolved in upward advecting porefluids driven by overpressures created through subduction-compaction or mineral dewatering reactions, and combinations of these (e.g., Kastner et al., 1991; Berndt, 2005). While geophysical investigations can identify the methane escape routes in the subsurface, the geochemical characteristics of the rising porefluids and gases are helpful in deciphering their origin and genesis (e.g., Hensen et al., 2004; Haffert et al., 2013).

For the southern Hikurangi margin, Crutchley et al. (2015) demonstrated that free gas zones of up to ~500 m thickness exist beneath the BSR, constituting the source for focused gas migration into the GHSZ. Fluid migration into the hydrate stability zone is inferred along conduits (Netzeband et al., 2010; Krabbenhoeft et al., 2013; Plaza-Faverola et al., 2014) and for sedimentary layers crossing the BSR (Crutchley et al., 2010, 2015).

This study presents a combination of geophysical data from Opouawe Bank and geochemical data from Opouawe Bank, Omakere Ridge, and Rock Garden, accretionary ridges at the Hikurangi margin offshore New Zealand. It shows that Opouawe Bank and the Hikurangi margin feature gas-only seeps and provide an insight on how gas migration may affect the gas hydrate system. The study discusses the shallow biogenic origin of the gas seeps as well as the migration pathways and focusing levels of the ascending gas, and thereby assesses if the seeps on Opouawe Bank can be used as a window to deep-seated processes of the subduction zone.

2.3. Geological setting

The Opouawe Bank (Fig. 2.1) is located at the narrowest part of the active Hikurangi margin, the southernmost expression of the Tonga-Kermadec-Hikurangi subduction zone. Dominated by the east–west oriented oblique subduction of the Pacific Plate underneath New Zealand's North Island, this part of the margin formed by accretion of 500 – 2,000 m thick marine sediments of the Hikurangi Plateau (Davy and Wood, 1994; Barnes et al., 2010). Subduction started approximately 24 Ma ago and, at present, the subduction rate is 49 mm/year at 37°S and declines southward to 40 mm/year at 42°S (DeMets et al., 1994; Collot et al., 1996; Barnes et al., 2010).

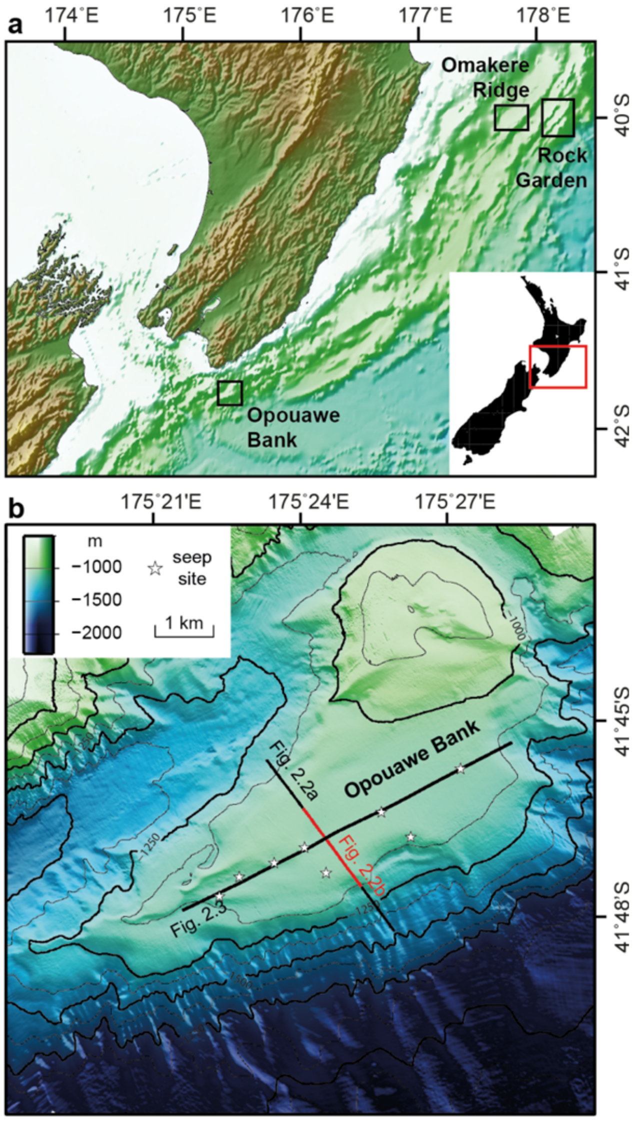

Figure 2.1: a Overview of the study area offshore New Zealand, with Opouawe Bank, Omakere Ridge, and Rock Garden. Geochemical analyses were carried out on material from a gravity corer and video-guided multiple corer in all three areas (for locations, see Table A.1 in the Appendix A.1). b Bathymetric map of Opouawe Bank, displaying the locations of the seep sites. The marked profiles are the 2D Nemesys line (Fig. 2.3), a segment of the 2D Pegasus line (Pegasus_19, Fig. 2.2a), and a section from the 3D P-Cable Cube (Fig. 2.2b).

Subduction has led to the formation of a series of anticlinal accretionary ridges parallel to the margin. Opouawe Bank, a SW–NE trending oval-shaped bathymetric high, is situated in 900 – 1,100 m water depth (Barnes et al., 2010) and associated with active methane seepage (Greinert et al., 2010). The NW flank of the ridge is characterized by translational landslide scars and the SE flank by gullies (Law et al., 2010). Opouawe Bank is separated from the continental slope by erosional canyons (Lewis et al., 1998), and is bordered in the south by the Hikurangi Trough (Barnes et al., 2010).

The tectonic structure of the area is dominated by three major sub-parallel fault systems (Barnes et al., 2010), which underlie the bathymetric highs on which the seeps are located. Eight seep sites (Fig. 2.1) were investigated in this study. They are located in the southwestern part of the Opouawe Bank, and have formed on the hanging wall of the Pahaua Fault (Barnes and Mercier de Lépinay, 1997). Six seep sites are aligned along the crest of the ridge, whereas two are located farther south on the ridge.

2.4. Materials and methods

2.4.1. Geophysical data

High-resolution seismic data (P-Cable system, Fig. 2.2) and 2D multi-channel seismic (MCS) data (Fig. 2.3) were acquired during the NEMESYS Project in 2011 from aboard the research vessel R/V SONNE (cruise SO214). The 2D MCS data were recorded using a 200 m long streamer with 128 channels and a group spacing of 1.5 m. The 3D data were recorded using a P-Cable system consisting of 16 streamers, each with 8 channels and a group distance of 1.5 m covering an area of 3 by 8 km. A single GI-gun with a volume of 210 cubic inches was operated in harmonic mode at a shot interval of 5 s. The MCS and P-Cable datasets both have a bandwidth of 50 – 300 Hz.

The main processing steps of the high-resolution 3D P-Cable seismic data included navigation correction for the source and the 16 streamers, trace editing, frequency filtering, and binning to a regular 3D grid with a cell size of 6.25 × 6.25 m. Due to insensitivity of moveout velocities, a water velocity stack was the input to 3D Kirchhoff time migration. The 3D velocity model for the migration was seafloor depth dependent interpolated and extrapolated from a 2D velocity profile crossing the 3D area.

The 2D single-streamer data processing included navigation processing, trace editing, frequency and velocity filtering, and a crooked line binning with cell size of 1.5 m. After water velocity stack, a post-Kirchhoff time migration with a representative velocity function below the seabed was applied.

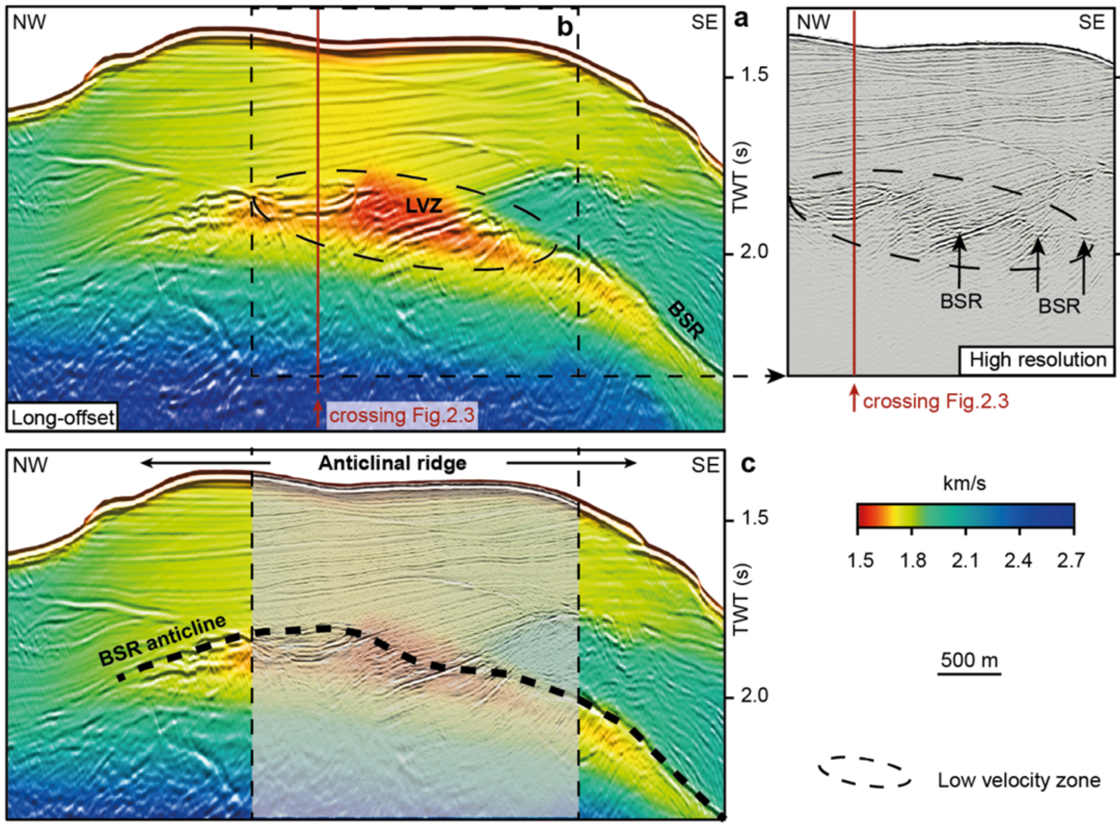

Figure 2.2: a Long-offset 2D Pegasus line and b corresponding cross section of the 3D Cube across the ridge. The plot in a displays the migrated depth section and the velocity distribution, with a low velocity zone (LVZ) underneath Opouawe Bank. Northwest dipping high-amplitude reflections occur in both sections at the location of the LVZ. c Merged section of a and b, illustrating the connection between the anticlinal structure of the ridge, the LVZ, and dipping high-amplitude reflections. The latter occur above and below the BSR.

Velocity information (Fig. 2.2a) was extracted from 2D MCS data originating from the PEGASUS survey (Ministry of Economic Development of New Zealand in 2009). A 31.5 km long northern segment of line Pegasus_19 (Plaza-Faverola et al., 2012) was reprocessed with special attention to the upper 1 km of sediments at Opouawe Bank for optimizing velocity analysis and pre-stack depth migration. The data were collected with a 10 km long streamer consisting of 800 channels with a group spacing of 12.5 m. Their dominant frequency of 25 Hz (zero-phase wavelet) leads to a vertical resolution (one-quarter of the dominant wavelength) of 8 – 10 m for a velocity range of 1,500 – 2,100 m/s. Frequency and wave number-domain filtering, deconvolution, and attenuation of seafloor and interbed multiples were applied using WesternGeco's Omega software during reprocessing at GEOMAR. The Kirchhoff pre-stack depth migration was done with Sirius by GX-Technology. The depth velocity field was iteratively determined from top to bottom by semblance and residual moveout analysis of pre-stack migrated common depth point (CDP) gathers. Five iterations resulted in a velocity error of less than 2.4%.

2.4.2. Geochemistry

Sediments of the Hikurangi margin from Opouawe Bank, Rock Garden, and Omakere Ridge (for geographic locations, see Table A.1 in Appendix A.1) were retrieved with a gravity corer and a video-guided multiple corer during expedition SO191 with RV SONNE in 2007. Porewater was extracted from ca. 2-cm-thick slices of sediment using a low-pressure squeezer (argon gas at 3 – 5 bar) at approximately in situ temperatures (4 – 8 °C) in the cold room onboard the research vessel. Upon squeezing, the porewater was filtered through 0.2 µm cellulose acetate Nuclepore filters and collected in recipient vessels. Onboard, the collected porewater samples were analyzed for their contents of dissolved Cl and SO_4 by ion chromatography, dissolved hydrogen sulfide by photometry, and dissolved total alkalinity by HCl titration (see Haffert et al., 2013 for analytical details and errors). Subsamples were taken for shore-based analyses of dissolved B, Li, Ca, Sr, and K concentrations using an atomic emission spectrometer, particulate organic carbon and calcium carbonate content using an element analyzer, and the isotopic composition of the water, i.e., $\delta^{18}O$ and δD, using an isotope ratio mass spectrometer (see Haffert et al., 2013 for analytical details and errors).

For later headspace methane and higher hydrocarbon gas analyses, 3 ml of wet sediment were collected and suspended in 3 ml of 10% KCl solution. Gas concentrations were measured by gas chromatography and the isotopic composition, i.e., $\delta^{13}C$, of the methane gas was determined by coupled gas chromatography and combustion isotope ratio mass spectrometry (Nuzzo et al., 2009). Undissociated gas hydrate samples from the cores were stored in liquid nitrogen containers, and the chemical and isotopic compositions were analyzed by the same procedures as the headspace gas samples.

2.5. Results

2.5.1. Seismics

Most seeps of Opouawe Bank align along the ridge crest. Their principal gas migration structure is discernible on the 2D NEMESYS line (Fig. 2.3) crossing along the ridge. Vertically it can be subdivided into three parts, starting at the base of the gas hydrate stability zone (BGHSZ), which is characterized by a band of high-amplitude reflections. This reflection band is terminated by blanking or by chaotic reflections with high-amplitude anomalies in the lower part of the migration structures. At the base and within the GHSZ, the gas migration structures converge into vertical conduits characterized by pull-ups in combination with reduced amplitudes or wipeout zones with weak reflections. The conduits cut vertically through the sub-horizontal strata, and some terminate at various stratigraphic levels. In the shallow sub-seafloor the conduits terminate as rounded structures at the seafloor with various amplitude anomalies, in some cases accompanied by gas-controlled doming of the seabed sediments (Koch et al., 2015).

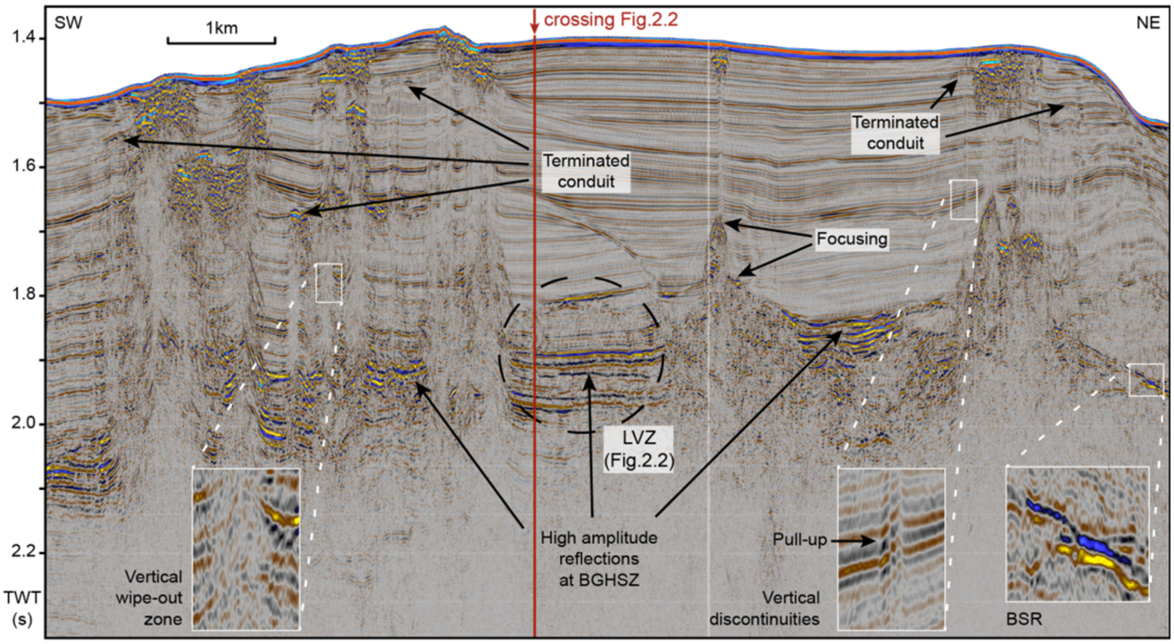

Figure 2.3: 2D Nemesys line along the Opouawe Bank crossing the seep structures aligned along the ridge. The vertical conduits display several focusing levels, and some conduits terminate within the sediments. The interface of Figs. 2.2 and 2.3 illustrates the spatial distribution of the high-amplitude reflections at the base of the gas hydrate stability zone (BGHSZ), where the LVZ is located.

The 2D Pegasus line (Fig. 2.2a) and a corresponding high-resolution seismic section (Fig. 2.2b) across the ridge connect seismic observations with the superimposed velocity field. At the center of Opouawe Bank, a low velocity zone (LVZ, Fig. 2.2a) is located underneath the anticline of the ridge, in about 320 mbsf (meter below seafloor). The seismic velocity decreases from about 1,715 m/s to about 1,575 m/s at the BGHSZ. Northwest dipping high-amplitude reflections crossing the BSR exist in the LVZ. These show the typical thickness variations and limited lateral extent of the seismic facies associated with contourite deposits. This is supported by the undulating interfaces approximately 200 ms below the seafloor, which are interpreted as boundaries between sediment waves. The crosscutting BSR in the Pegasus line has reversed polarity on both flanks of the anticline, whereas it is not a proper reflector at the apex. Apart from the crosscutting BSR, there are no other crosscutting seismic reflectors, suggesting that out-of-plane reflections do not significantly affect the seismic imaging. Patches of high-amplitude reflections and patches of reduced reflectivity occur also at the peak of the LVZ on the Pegasus line (Fig. 2.2a). Beneath the LVZ, the Pegasus line displays subdued reflections and low seismic resolution as consequence of low frequencies.

2.5.2. Geochemistry

In samples retrieved from different seeps at Opouawe Bank, there are only slight variations in the isotopic composition of porewater, dissolved chloride concentrations, carbon isotopic composition of the dissolved methane, and the gas composition (Fig. 2.4). This is in line with data from other seep areas of the Hikurangi margin, such as Rock Garden and Omakere Ridge (cf. Fig. 2.4). Porewater isotope signatures and chloride concentrations show typical seawater values (δD of -4 to -2‰ SMOW, $\delta^{18}O$ of -0.5 to +0.5‰ VPDB, and Cl of 540 to -570 mM), with possibly small influences of near-surface formation of gas hydrates and the corresponding artifact caused by the dissociation of gas hydrates during core retrieval (Fig. 2.4a, b; for a detailed discussion on this issue, see Haeckel et al., 2004). Boron and lithium concentrations have similar trends, i.e., B of 0.3 to -0.5 mM and Li of 15 to -25 μM (see Fig. A.2 in Appendix A.1).

The dissolved seep gas is composed primarily of methane (C_1/C_{2+} ratio >500), and the methane is isotopically enriched in light ^{12}C, thus carrying a $\delta^{13}C$ of -60 to -80‰ VPDB (Fig. 2.4c). The gas bound in near-surface gas hydrates at the Takahe seep is naturally even more enriched in methane (C_1/C_{2+} of 11,000 to -23,000). This methane has isotopic signatures of $\delta^{13}C$ of -65 to -66‰ VPDB and $\delta^{2}H$ of -126 to -145‰ VSMOW (Fig. 2.4d).

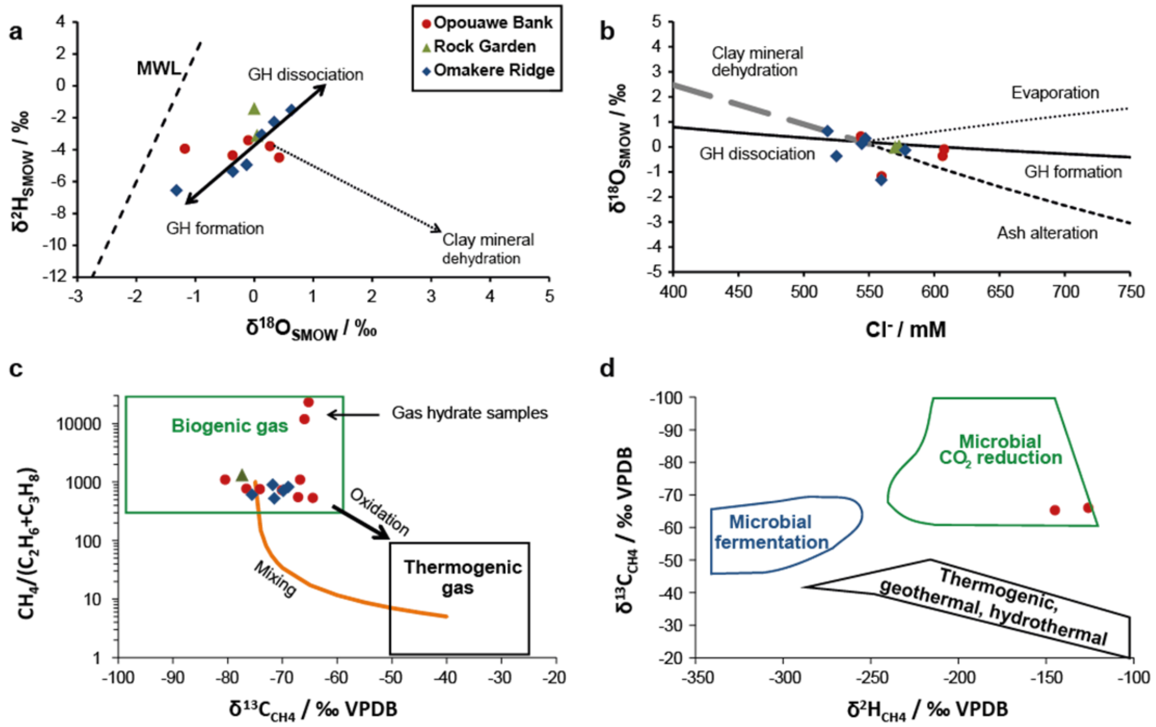

Figure 2.4: Geochemical characterization of the porefluid and gas composition as well as gas hydrates in surface sediments of cold seeps at the Hikurangi margin (Opouawe Bank, Rock Garden, and Omakere Ridge). a Porewater isotopic composition and typical trends from gas hydrate formation and dissociation (solid line), clay mineral dehydration (dotted line), and meteoric water mixing (dashed line; MWL meteoric water line). b Porewater oxygen isotopes versus chloride concentration, and influences from typical diagenetic processes. c Bernard plot indicating the biogenic or thermogenic origin of the gas. For comparison, the theoretical mixing line between biogenic and thermogenic methane as well as the trend for oxidation are indicated. d Isotopic composition of methane in the recovered gas hydrate pieces from the Takahe seep, and process-related fields (according to Whiticar et al. 1986).

2.6. Discussion

2.6.1. Gas focusing

The strong decrease of seismic velocities at the LVZ from 1,715 to 1,575 m/s, which coincides precisely with the base of the GHSZ, suggests gas accumulation in sediments of Opouawe Bank at about 320 mbsf. The shape of the LVZ at the BGHSZ indicates trapping of upward migrating gas under the anticlinal structure of the ridge, which forms a trap due to the reduction of permeability in the hydrate sediments (Bünz et al., 2003). This entrapment has been observed for many gas hydrate provinces (Bünz et al., 2003; Crutchley et al., 2013). The coincidence of the LVZ with the BSR argues against other possible interpretations such as underconsolidation or lithological variations, as these would have to be located at this depth and dipping strata coincidentally. Decreased seismic velocities underneath the oblique BSR and the inclined base of the LVZ (Fig. 2.2) indicate ascending gas toward the LVZ at the southeastern flank of the ridge. This suggests that free gas originates from deeper parts of the sediments, and is not generated at the depth of the LVZ.

The LVZ extends upward beyond the BSR depth (Fig. 2.2). It corresponds to a band of near-horizontal high-amplitude reflections in between the seep structures (Fig. 2.3) and where the BSR is highly attenuated or inexistent (Fig. 2.2). The observed BSR gap corresponds with an area where the LVZ reaches up into the GHSZ, suggesting that there is no sharp impedance contrast across the BGHSZ, as gas is migrating upward into the GHSZ. Similar observations have been documented in the proximity of thrust ridges at the southern Hikurangi margin (Crutchley et al., 2015) and for the Gulf of Mexico (Boswell et al., 2012).

The formation of migration pathways under the ridge structure and the alignment of most seep sites along the ridge crest are most likely linked to anticlinal focusing (Johnson et al., 2003). As the Opouawe Bank is an anticlinal structure (Barnes et al., 2010; Law et al., 2010), the BSR also adopts an anticlinal shape. At the BGHSZ the gas hydrate constitutes a permeability contrast (Hornbach et al., 2004). Bünz et al. (2003) showed that the decreased permeability enables gas accumulation, which probably generates overpressure and particular high overpressure below an anticlinal BSR leads to the formation of fluid migration pathways. This overpressure is at least partly due to buoyancy forces, but can be augmented by dynamic flow processes (Crutchley et al., 2014). Thus, subsequent fracturing may lead to focusing of gas migration into the GHSZ.

The 2D NEMESYS line (Fig. 2.3) reveals several focusing levels for the ascending gas. This is reflected by the seismically distinguishable parts of the gas migration structures, first at the BGHSZ where gas migration from the LVZ into the GHSZ is accompanied by blanking or by chaotic reflections. Successive focusing occurs within the GHSZ, which in some parts is visible as a cone-like shape with high-amplitude anomalies, to vertical narrow conduits characterized by pull-ups in combination with reduced amplitudes or wipe-out zones with weak, deteriorated reflections.

The conduits exhibit no specific stratigraphic level at which their tops cluster, which indicates a long record of gas migration through Opouawe Bank. This is supported by different evolutionary stages of the gas migration structures observed in the upper 100 m in the shallow sediments (Koch et al., 2015) and the different phases of intensified seep activity, which Liebetrau et al. (2010) inferred from geochemical analysis of authigenic carbonate samples.

2.6.2. Gas source

At the Hikurangi margin, porewater oxygen and hydrogen isotopic signatures from subsurface sediment cores (uppermost 6 m) concomitantly show a positive correlation, i.e., δD values increase with increasing $\delta^{18}O$ values (Fig. 2.4a), whereas $\delta^{18}O$ values are negatively correlated with Cl concentrations (Fig. 2.4b). Both trends are indicative for currently active in situ gas hydrate formation in the sediments, as well as dissociation of gas hydrates due to depressurization of the retrieved cores (see Haeckel et al., 2004 for detailed discussion). This is, however, the complete opposite of what is to be expected if fluid advection would be driven by clay mineral dewatering reactions, i.e., smectite-illite transformation occurring at temperatures of 60–190 °C (cf. dotted arrow in Fig. 2.4a and dashed line in Fig. 2.4b). In accordance with this observation, porewaters are not enriched in boron and lithium, as would be the case if upward fluid flow driven by smectite-illite dewatering took place (e.g., Hensen et al., 2004, 2007; Haffert et al., 2013). Instead, Li and B concentrations decrease downcore (see Fig. A.1 in Appendix A.1), which is a typical result from uptake during ash alteration (e.g., Kastner and Rudnicki, 2004; Scholz et al., 2010).

In line with the above observations, also the seeping gas does not show any indications for a deep origin, i.e., a thermogenic imprint caused by temperatures above 80 °C. The expelled gas as well as the one bound in near-surface gas hydrates is composed primarily of methane, generated from organic matter degradation via metabolic CO_2 reduction (Fig. 2.4c, d). Considering the observed margin-wide heat flow (Henrys et al., 2003), the source depth of the methane is at maximum 1,500-2,100 m below the seafloor and may shoal below seeps, such as Takahe, to 600 m sediment depth due to increased heat flow (Schwalenberg et al., 2010a). Thus, the methane at Opouawe Bank and in the other investigated seep areas at the Hikurangi margin is largely generated below the GHSZ (Uruski and Bland, 2011; Kroeger et al., 2015), as is expected (Wallmann et al., 2006, 2012).

In addition, the 2D seismic lines do not reveal possible focused gas migration pathways in greater depths underneath the BGHSZ (Fig. 2.3) and underneath the LVZ (Fig. 2.2b). However, some of the conduit structures might be of a deeper origin than the BSR level. Coherent reflections are not present below the LVZ, and thus it is not clear if there are focused fluid migration pathways. Therefore, the origin of the upward migrating gas below the LVZ cannot be determined in the seismic data at Opouawe Bank.

Thus, the present study does not document any evidence for fluid seepage through the deep thrust faults underlying Opouawe Bank, as hypothesized by Plaza-Faverola et al. (2012). Migration of thermogenic gas is postulated along the décollement into the sediments of the Pegasus Basin (Kroeger et al., 2015), but does not appear to have a connection to the seeps further upslope, such as Opouawe Bank.

Further evidence that the methane transport is dominated by gas migration through the surface sediments is provided by the ubiquitous occurrence of S-shaped profiles of dissolved porewater constituents (Schwalenberg et al., 2010a, 2010b). Boudreau et al. (2005) have established that gas bubbles rise through soft, muddy surface sediments by linear elastic fracture mechanics, thereby creating tubular rise paths, and Haeckel et al. (2007) have demonstrated that this results in irrigation-like mixing of bottom water solute concentrations several meters into the sediments. In contrast, upward porewater advection would lead to concave-shaped solute profiles.

2.7. Conclusions

This study demonstrates the strength of linking geochemical and geophysical data in order to discriminate between gas ascent and the upward advection of water. The analyses clearly show that the cold vents at Opouawe Bank (as well as at the Hikurangi margin in general) are fueled by the seepage of biogenic methane gas. Bubble-induced irrigation-type porewater profiles, strongly enhanced methane fluxes, and the absence of any signal of mineral dewatering reactions provide the main geochemical evidence. Hence, seepage at Opouawe Bank appears not to be driven by the compaction-subduction process, but by upward migration of microbial gas. The source depth of the methane is at maximum 1,500–2,100 mbsf, considering the average margin-wide heat flow, which does not allow microbial gas production at greater depth.

The combination of MCS data and velocity analysis shows that gas migration through Opouawe Bank is a continuous process. Trapping of upward migrating gas under the anticlinal structure of the bank and the alignment of most of the seeps along the ridge crest, document anticlinal focusing of the ascending gas. In addition to gas migration along focused pathways through the gas hydrate stability zone, high-amplitude reflection segments indicate migration of gas along layers crossing the bottom simulating reflection.

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3. Elongated fluid flow structures

Stress control on gas migration at Opouawe Bank, New Zealand

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This chapter will be submitted to a peer-reviewed journal.

3.1. Abstract

Focused vertical fluid migration through marine sediments commonly manifests as pipes or chimneys on seismic images. Understanding the mechanics of these fluid flow systems can yield insights into tectonic deformation processes and the distribution of carbon in the subsurface. Here, we present high-resolution 3D seismic data from Opouawe Bank, an accretionary ridge on the Hikurangi subduction margin, offshore New Zealand, providing evidence for exceptional gas migration pathways. The spatial analysis revealed parallel and horizontally elongated structures perpendicular to the strike of the ridge instead of the concentric cross-section of typical pipe structures. Only in about 75 to 100 m below the seafloor a transition from elongated structures at depth to concentric pipe structures above takes place and the pipe structures continue to the seafloor where they terminate at seeps. Gas migration along the extensional fractures causes easily identifiable seismic anomalies providing robust insight into the stress state within the continental margin. Compressive stress leads to the formation of Opouawe Bank parallel to the margin and the secondary longitudinal extension of the ridge contributes to the upward gas migration along extensional fractures. With declining differential stress in the shallow sediment, a transition to pipe-like structures is possible.

3.2. Introduction

The relationship between gas migration, gas hydrates, and seafloor seepage has implications for the understanding of subduction zone processes and the interaction between the global carbon reservoir and seabed ecology. Comprehending the mechanics of fluid migration at subduction zones is important as fluid exerts control on interplate seismogenesis (Ranero et al., 2008). Fluid migration also has a significant impact on the distribution of carbon in the subsurface and the amount of carbon leaking from the seafloor (Berndt, 2005) and thereby cold seep systems (Hovland et al. 2002; MacDonald et al. 2002; Sibuet und Olu, 2003). Gas seepage from marine sediments comprises a differentiated system from the source and the plumbing structures, to seep structures at the seabed (e.g. Talukder, 2012; Andresen, 2012). At active continental margins the migration of fluids or gas can provide insights into processes of tectonic deformation, the reduction of porosity and compaction of the sedimentary sequences (Kvenvolden, 1993; Kopf, 2002; Suess, 2014).

The plumbing or hydrocarbon leakage systems (Cartwright et al., 2007; Løseth et al., 2009; Andresen, 2012) from the reservoir to the seabed are associated with structural migration along faults and fractures or stratigraphically-controlled migration (Talukder, 2012). Seismic reflection imaging is an ideal way to investigate the nature of gas-charged fluid migration beneath the seafloor, since focused flow can have profound effects on the reflectivity of sediments. Vertical fluid flow conduits crosscutting the sedimentary strata are often called seismic pipes or chimneys (Karstens and Berndt, 2015) and usually appear as columnar zones of seismic blanking, turbidity or reduced amplitudes, caused by absorption and scattering of acoustic energy by the gas charged sediments (e.g. Judd and Hovland, 1992; Riedel et al., 2002; Wood et al., 2002; Gay et al., 2007; Løseth et al., 2009, Husthoft et al., 2010). Migrating fluids that are expelled at the seabed into the water column are often linked to various seafloor features, such as seep fauna, carbonate precipitates, mud volcanos, pockmarks, mounds and seabed domes (Hovland and Judd, 1988 and references therein).

On the Hikurangi margin, offshore New Zealand's North Island several areas with multiple seep sites are present on the crests of anticlinal ridges situated in 700 - 1200 m water depth(Greinert et al., 2010; Barnes et al., 2010). The seeps are located within the gas hydrate stability zone (GHSZ) and both structurally and stratigraphically controlled fluid migration systems are sustaining the seep sites on these thrust-folded accretionary ridges (Barnes et al., 2010; Crutchley et al., 2010; Krabbenhoeft et al., 2013). At active margins in the accretionary wedge, overpressure is in most cases not sufficient to induce hydrofracturing and fluid flow is mainly initiated by external factors and tectonic stress (Talukder, 2012).

In this paper we investigate the nature of gas-charged fluid flow beneath Opouawe Bank, an accretionary ridge at the Hikurangi Margin, offshore New Zealand, which is host to 13 seep sites (Greinert et al., 2010). This distribution makes it one of the most densely populated regions of methane seepage known offshore New Zealand. Since these seep sites sustain diverse biological communities and might point to concentrated gas hydrate deposits at depth, we seek to understand which geological conditions favor such focused fluid flow. Using 3D seismic data, our objective is to image and map out the specific structures that allow such prolific gas migration through the gas hydrate layer. Our results will give insight into the local stress conditions beneath the ridge and how they relate to the mechanics of gas migration.

3.3. Geological setting

The 25 Myr old active Hikurangi Margin of eastern North Island, New Zealand, is the southernmost expression of the Tonga-Kermadec-Hikurangi subduction zone, where westward subduction accommodates oblique convergence between the Pacific Plate and the Australian Plate (Barnes et al., 2010). At present the subduction rate is 49 mm/yr at 37°S and declines southwards to 40 mm/yr at 42°S. Southwest of 42°S, strike slip motion begins to dominate (De Mets et al., 1994; Collot et al., 1996; Beavan et al., 2002; Barnes et al., 2010).

The subduction is increasingly more oblique southwards with an angle of 60° in the north and 20° in the south, as a result of variation in the plate boundary orientation and the direction of the relative motion between the plates (Wallace et al., 2012). The margin-normal component of the plate motion decreases southwards and is about 20 mm/yr at the Wairarapa study area (Fig. 3.1), located at the narrowest part of the margin. Most of the margin-normal component is accommodated by the subduction thrust (Barnes et al., 1997; Wallace et al., 2012) and the margin-parallel component, with about 30 mm/yr in the southern North Island, constitutes strike-slip faulting in the upper plate and forearc block rotation (Beanland and Haines, 1998; Wallace et al., 2004; Wallace et al, 2012).

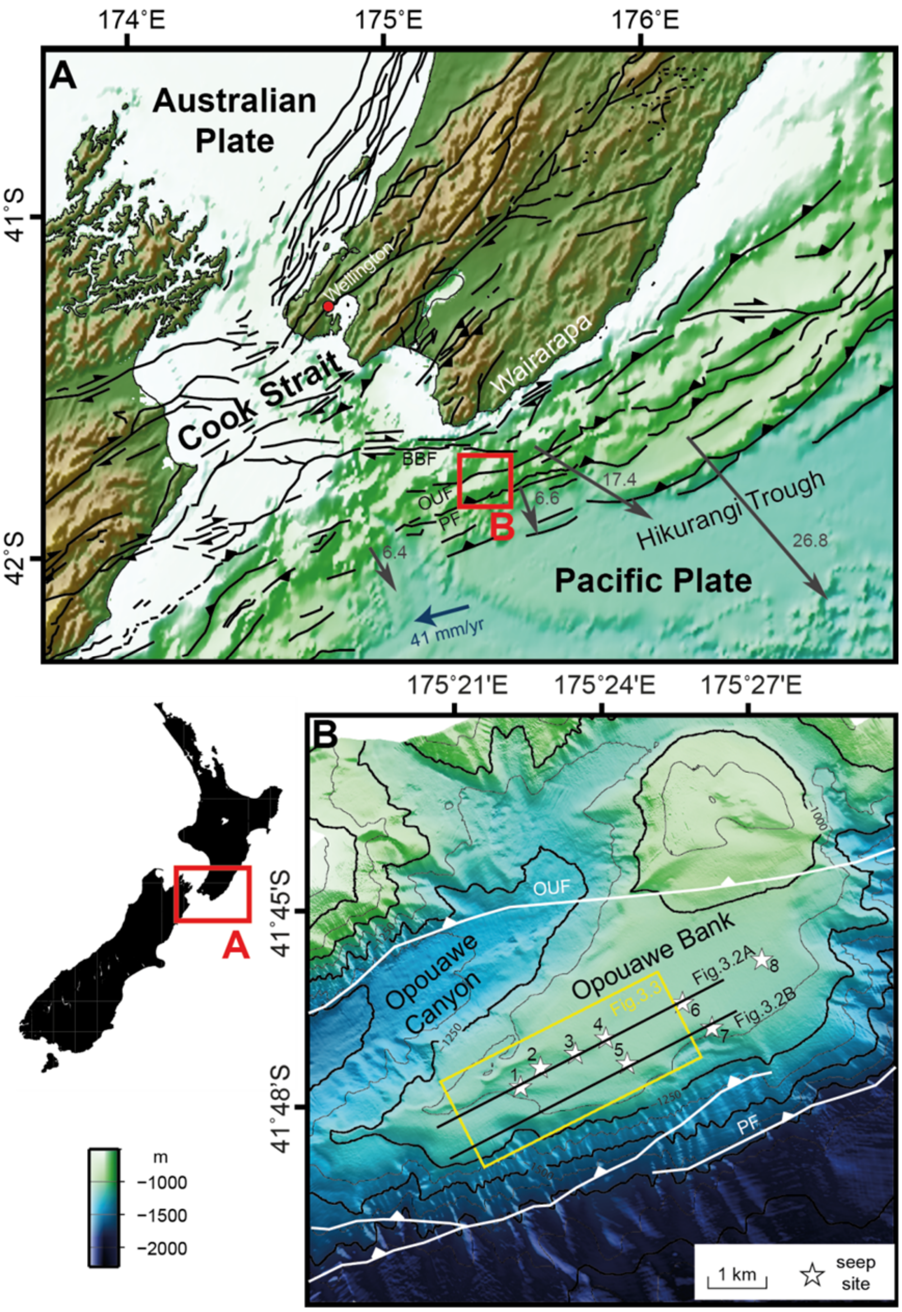

Figure 3.1: A) Bathymetric map showing the location of Opouawe Bank at the Hikurangi margin, offshore New Zealand. Geological structures are from Wallace et al. (2010). Blue arrow shows the relative plate motion vector of the Australian and Pacific plates at the southern North Island, with most of the margin normal component of the plate motion occurring on the subduction thrust and the margin parallel component as a combination of strik-slipe faulting and forearc rotation (Wallace et al., 2010 an references therein). Gray arrows display the modeled relative motion (in mm/yr) between tectonic block boundaries in the east and the Pacific plate (Wallace, 2012). BBF = Boo Boo Fault; OUF = Opouawe-Uruti Fault; PF = Pahaua Fault. (B) Location of seep sites (white stars) on Opouawe Bank investigated in this study 1 = Piwkawaka; 2 = Riroriro; 3 = Pukeko; 4 = North Tower; 5 = South Tower; 6 = Takahe; 7 = Takapu; 8= Piwakawaka. The yellow rectangle displays the outline of the 3D seismic volume (Fig. 3.3) and the black lines indicate the location of the 2D MCS data (Fig. 3.2).

The accretionary wedge narrows from about 80 km in the central part of the margin to 13 km at the southern end and thus displays abundant frontal accretion under very oblique convergence. The accretion has led to the formation of right-stepping, thrust-faulted and folded anticlinal ridges parallel to the margin that stand up to 1 km above the surrounding seafloor (Barnes et al., 1997). Opouawe Bank, a SW-NE trending oval-shaped bathymetric high, is one of these ridges in the Wairarapa area, culminating in about 1000 m water depth (Barnes et al., 2010). Separated from the continental slope by erosive canyons (Lewis, et al. 1998) and delimited in the south by the Hikurangi Trough (Barnes et al., 2010), the SE flank of Opouawe Bank is characterized by gullies and the NW flank by translational landslide scars (Law et al., 2010). The most recent sediments on the ridge top are hemipelagic mud and turbidity current overspill deposits (Lewis et al., 1998). The tectonic structure of the Wairarapa area is dominated by three major sub-parallel fault systems; these are, from north to south, the strike-slip Boo Boo Fault, and the Opouawe-Uruti and Pahaua thrust faults (Barnes and Mercier de Lépinay, 1997; Barnes et al., 2010).

This study investigates the subsurface structures of eight seep sites (Fig. 3.1), located on the south-western part of the ridge on the hanging wall of Pahaua Fault.The seep sites are located within the GHSZ and multi-channel seismic (MCS) data show a Bottom Simulating Reflector (BSR) underlying the flanks in the southwest and northeast of Opouawe Bank from 0.7 s TWT (two way traveltime) to 0.35 s TWT beneath the crest (Netzeband et al., 2010; Krabbenhoeft et al., 2013). Acoustic manifestations of the subsurface gas migration structures from MCS data have been reported by a number of authors (e.g. Law et al., 2010; Netzeband et al., 2010; Plaza-Faverola et al., 2012; Krabbenhoeft et al. 2013; Koch et al., 2015); the general mechanism of methane migration through the GHSZ at Opouawe Bank was described as structurally controlled (Law et al., 2010; Krabbenhoeft et al., 2013) and Law et al. (2010) concluded that fluid venting is promoted by geological features that include extensional faults and fracture networks and particular stratigraphic pathways. The source depth of the biogenic methane that feeds the seep sites is about 1500-2100 meters below the seafloor (Koch et al., 2016) and the upward migration of methane is influenced by anticlinal focusing (Law et al., 2010). In the upper 100 m below the seafloor, different evolutionary stages in individual gas migration structures and gas-controlled seafloor doming are reported (Koch et al., 2015).

3.4. Methods

3D P-cable seismic data, 2D MCS data and parametric echosounder (Parasound) data were acquired during the Nemesys Project in 2011 aboard R/V SONNE (cruise SO214). The 3D data covering an area of 3 by 8 km were recorded using GEOMAR's P-cable system with 16 parallel towed streamers. Each streamer consisted of 8 channels at a group spacing of 1.5 m. The 2D MCS data were recorded using a 200 m-long streamer with 128 channels and a group spacing of 1.5 m. A single GI-gun with 210 cubic inches volume was operated in harmonic mode with a shot interval of 5 s. The 2D and 3D seismic data have a mean frequency spectrum of 50-300 Hz after processing.

The main processing steps of the high resolution 3D P-Cable seismic data included navigation correction for the source and the 16 streamers trace editing, deghosting, frequency filtering and binning to a regular 3D grid with a nominal cell size of 3.125 x 3.125 m. Due to insensitivity of moveout velocities a water velocity stack was the input to 3D Kirchhoff time migration. The 3D velocity model for the migration was seafloor depth dependent interpolated and extrapolated from a 2D velocity profile crossing the 3D area. The 2D single-streamer data processing included navigation processing, trace editing, frequency and velocity filtering, and a crooked line binning with cell size of 1.5 m. After water velocity stack a post Kirchhoff time migration with a representative velocity function below the sea bed was applied.

Parasound profiles were collected in tandem with the 3D seismic data over an area of 3 × 8 km with an average line spacing of 50 m and additional single 2D profiles. The sub-seafloor was imaged at 4 kHz, with the secondary low-frequency component of the system, providing a subbottom penetration of 100 m and more, with a vertical resolution at decimeter scale. To identify free gas emission in the water column, the primary high-frequency component with 18 kHz was used.

To aid seismic interpretation, we calculated similarity volumes from the 3D seismic data with OpendTect software (dGB Earth Science; http://www.opendtect.org/). The similarity is the coherence of each trace over a defined window length (28 ms and 40 ms) and was computed by comparing the data in a time-window, with data in the equivalent windows of 3 neighboring traces. Hence, similarity highlights discontinuities in the data, such as faults, fractures and unconformities (Chopra and Marfurt, 2007). We chose a window-length of 40 ms to examine the spatial structure of the elongated structures, as they cut vertically through the sedimentary strata and a window-length of 28 ms to investigate their transition to pipe like structures in the shallow subsurface. Elongated zones of low similarity were picked on time-slices of the similarity volume of the 3D seismic data with Kingdom Suite (IHS; www.ihs.com/products/kingdom-seismic-geological-interpretation-software.html). We used the Generic Mapping Tools (GMT; Wessel and Smith, 1998) to generate rose diagrams from the azimuths and lengths of the fractures.

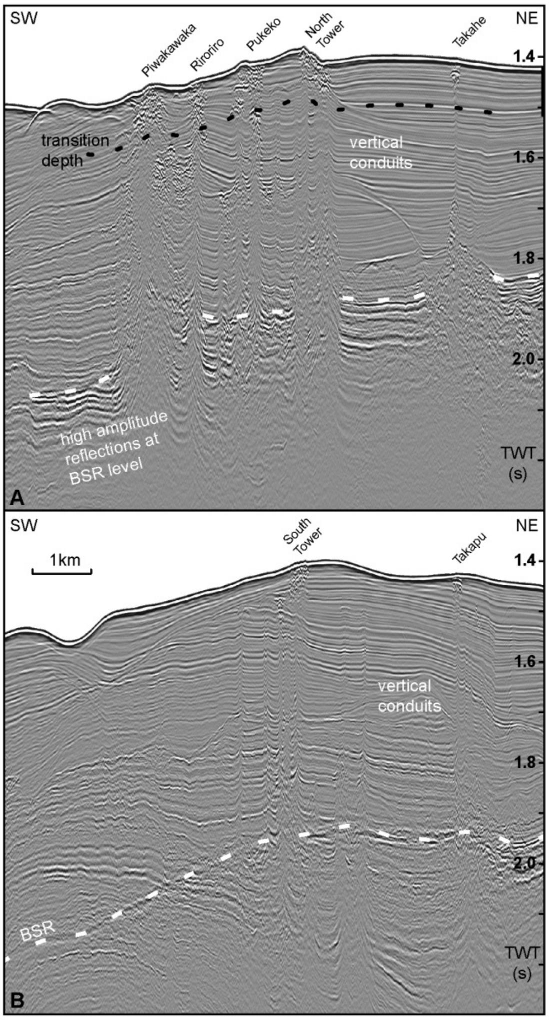

Figure 3.2: A) Northern and B) southern 2D seismic line along the main seep sites on Opouawe Bank, displaying the gas migration pathways through the accretionary ridge. These pathways appear as vertical conduits of limited extend on the 2D seismic. However, Figure 3.3 displays their spatial structure to be elongated across the ridge. Marked as dashed black line is the depth, where the gas migration structures transform to pipe-like structures.

3.5. Results

The 2D seismic data (Fig. 3.2) along the ridge display vertical gas migration pathways that appear as seismic wipe-out zones or narrow conduits. The conduits are 50 to 85 m in width along the ridge and they are characterized by reduced amplitudes. The wipe-out zones are about 90 – 220 m wide, with weak deteriorated reflections. Both structures display vertical discontinuities accompanied by reflection pull-ups, in the mostly sub-horizontal sediment strata of the ridge. The similarity time-slices (Fig. 3.3) of the 3D survey reveal that these vertical conduits and wipe-out zones are actually elongated structures across the ridge. In the upper 75 - 100 m below the seafloor a transition to a rounded shape, as observed at the seafloor, occurs. At some sites these gas migration structures are connected to gas-controlled doming of the seabed sediments (Koch et al., 2015).

3.5.1. Elongated seismic anomalies

Figure 3.3 visualizes with five similarity-slices and the similarity map of the seafloor reflection, the spatial relationship between the western seep sites and the gas migration structures within the sediment. The elongated structures (Fig. 3.3) are located underneath the seep sites on the hanging wall of the Pahua thrust fault. They are more or less parallel to each other and perpendicular to extension to the SW-NE trend of the ridge. Some of the elongated structures terminate at various depths beneath the seafloor (Fig. 3.2).

The number of elongated structures that contribute to the seep sites varies for the different sites. It is not clear where the root of these structures is located at depth (Fig. 3.2). Some of them extend beneath the BGHS, as visible on Figure 3.2B. However, underneath the seep structures at the crest of the anticlinal ridge (Fig. 3.2A) we cannot resolve the structures underneath the BGHS due to accumulation of ascending methane (see Koch et al., 2016, for a detailed discussion on this issue).

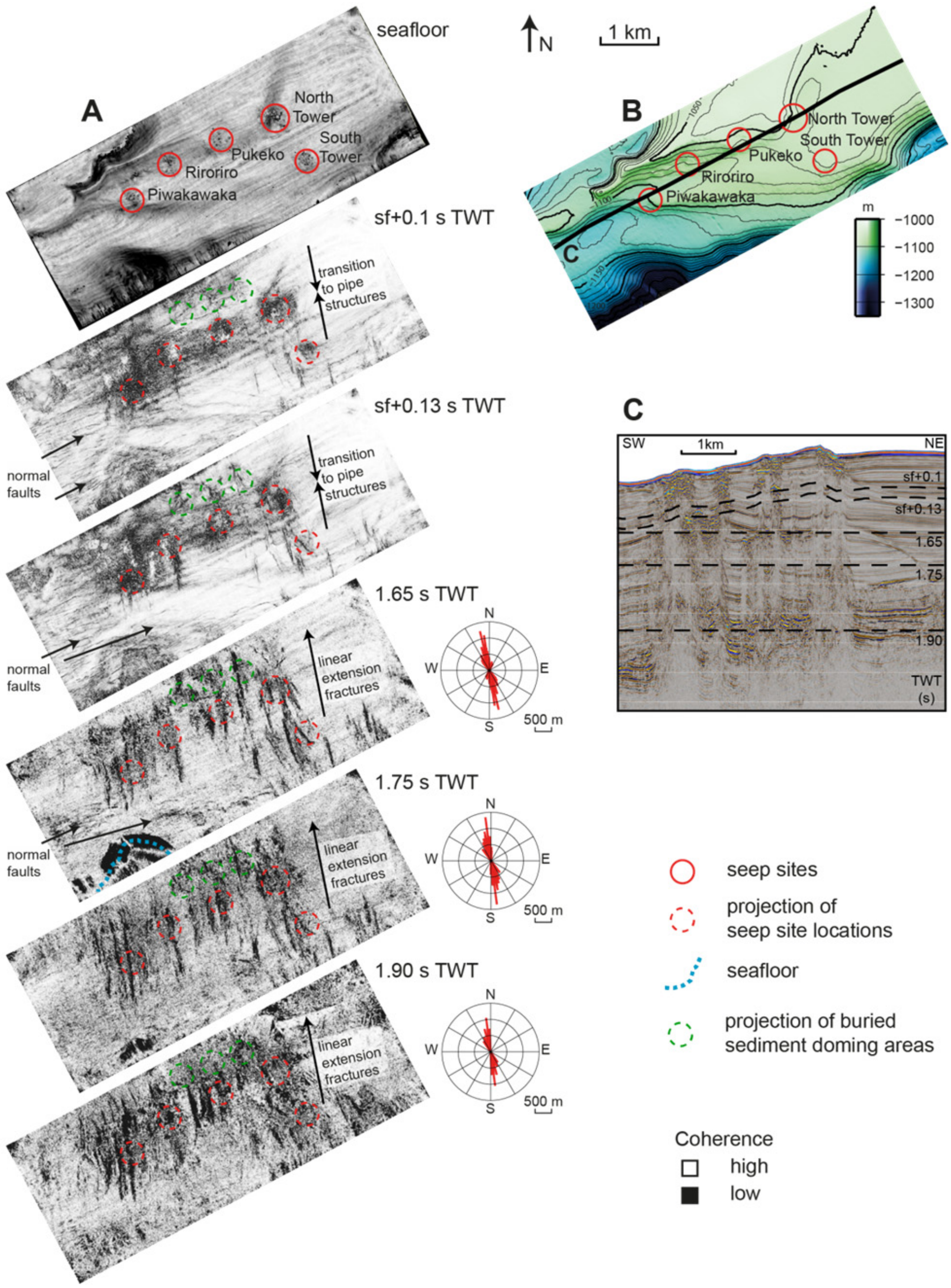

Figure 3.3: A) displays the similarity maps (high coherence – white, low coherence - black) extracted from the seafloor, horizons parallel to the seafloor (sf+0.1, sf+0.13 s TWT), and as time-slices (1.65, 175, 1,90 s TWT). B) shows the locations of the seep sits at the seafloor, on a bathymetric map (complementary to the seafloor similarity of A). C) displays the position of the similarity maps (A) on a 2D seismic profile. The similarity discloses the spatial structure of the gas migration through Opouawe Bank. Gas migrates along linear extensional fracture systems (1.65, 175, 1,90 s TWT; rose diagrams indicate the orientation and extent) and in the shallow sediment a transformation to pipe-like structures takes place (sf+0.1, sf+0.13 s TWT).

Within the GHSZ the elongated structures have a larger extent perpendicular to the ridge (Fig. 3.3) and a shorter extent along the ridge (Fig. 3.2). The lateral extent below 75 – 100 mbsf, for the structures beneath the four westernmost seep sites Piwakawaka, Riroriro, Pukeko, and North Tower (Fig. 3.3), can be traced perpendicular to the ridge over distances as great as 1000 - 1300 m. In general, the time-slices from 1.65 s TWT to 1.9 s TWT illustrate an almost constant length. At the SW flank of the ridge normal faulting occurs (Fig. 3.3, time-slice 1.65 s TWT). The faults strike parallel to the strike of the ridge and dip at an angle of 20°- 35° northwestwards into the ridge. The normal faults are not connected with the elongated structures and not associated with seepage.

The elongated structures at South Tower have an extent of about 500 m and are connected with the structures beneath North Tower. To the north, the elongated structures connect to buried gas migration structures (Fig. 3.3; Koch et al., 2015). The eastern three sites Takapu, Takahe, and Papango (Fig. 3.2) lie outside the 3D volume and were tracked on Parasound lines, which have a penetration of 100 m and more. In spite of the limited data coverage outside of the 3D grid, we were able to trace the extent of the elongated structures as being up to 300 - 500 m. The elongated structures beneath Takahe and Takapu appear to be connected with each other, like the structures of North and South Tower.

3.5.2. Transition from elongated to rounded gas migration structures

Approximately 75 – 100 mbsf (Fig. 3.3), the elongated structures become shorter and converge into almost rounded structures. This transformation is visible at similarity-slices with constant shifts parallel to the seafloor (sf+0.1 and sf+0.13 s TWT, Fig. 3.3). The structural change of the pathways does not take place at a specific stratigraphic horizon; rather it appears to be controlled by the sediment depth beneath the seafloor.

The shallow gas migration structures display indicators for gas accumulation (Koch et al., 2015) and involve gas trapping beneath relatively low-permeability horizons, overpressure accumulation, sediment doming and the subsequent development of methane seep sites. At seafloor the seep sites have a more or less rounded shape, with diameters between 250 and 500 m. The similarity time-slice of the seafloor (Fig. 3.3) displays the surface texture at the seeps. The relatively incoherent nature of the seafloor reflection at the seep sites is the result of carbonate precipitates, which have been described from sidescan sonar data (Dumke et al., 2014; Klaucke et al., 2010).

3.6. Discussion

3.6.1. The nature of the elongated seismic anomalies

The presence of gas in sediments affects the impedance contrast and causes seismic anomalies, as acoustic energy is scattered, reflected and attenuated (Sheriff and Geldart, 1995). Vertical anomalies, such as seismic wipe-out zones and narrow conduits are often associated with reduced amplitudes caused by attenuation of the acoustic signal and indicate upward gas-charged fluid migration, or even destruction of the original sediment layering (Hovland and Judd, 1988; Schroot et al., 2005, Løseth et al., 2009). The absorption and scattering of acoustic energy by gas can also cause deteriorated reflections or acoustic turbidity (e.g. Judd and Hovland, 1992; Yuan et al., 1992) and is sometimes observed beneath high-amplitude reflections caused by free gas accumulations. Therefore, we conclude that the seismic wipe-out zones with low amplitudes and limited lateral extents (Fig. 3.2), and narrow anomalies with reflection pull-ups, high amplitude anomalies, and phase reversals indicate the ascent of gas through Opouawe Bank.

This is supported by the observation of gas seepage at many of the vent sites (Greinert et al., 2010) and evidence for shallow gas accumulation and migration structures in the upper 100 m below the seafloor (Koch et al., 2015; Klaucke et al., 2010). Hence, the gas feeding these structures migrates along the vertical pathways through the GHSZ on Opouawe Bank. This process was previously described to take place along vertical chimneys beneath the seep sites (Netzeband et al., 2010; Krabbenhoeft et al., 2013). Although these studies lacked the spatial information provided by the 3D survey, Krabbenhoeft et al. (2013) showed that chimney structures are offset with respect to the seeps observed at the seafloor.

Diverse structures are known to act as focused migration pathways for fluids and gas beneath cold seeps; such structures are referred to as hydrocarbon leakage or plumbing systems (Løseth et al., 2009; Talukder, 2012; Andresen, 2012). Typical manifestations of vertical gas migration on seismic images are pipes, chimneys and polygonal fault systems. The similarity time-slices of the 3D cube (Fig. 3.3) clearly show that the gas migration structures in Opouawe Bank are in fact elongated with a larger extent perpendicular to the ridges extent in the interval between 100 ms below the seafloor and the BGHS (Fig. 3.4). Below the BGHS, structures cannot be resolved due to gas accumulation (Koch et al., 2106) and the limited penetration of the high-resolution 3D seismic system. It seems gas migration through Opouawe Bank occurs along parallel elongated pathways, which is highly unusual as such vertical fluid migration structures are mostly concentric or elliptic (e.g. Husthoft et al., 2010). However, elongated pathways have been described for a conjugate Riedel shear zone at Omakere ridge (Plaza-Faverola et al., 2014). We propose that elongation of the fluid migration structures is the result of an anisotropic stress regime within the anticlinal ridge below that the shape of the fluid migration structures holds information on the stress pattern.

3.6.2. Implications for the stress regime

On the margin scale, the relative motion between the Pacific and Australian plate is oriented approximately WSW (Fig. 3.1, 3.4) at the southern end of the Hikurangi subduction zone, whereas the margin-parallel component is compensated by strike-slip faulting in the upper plate and forearc block rotation (Beanland and Haines, 1998; Wallace et al., 2004; Wallace et al, 2012). The margin-normal component is mostly accommodated by the subduction thrust (Barnes et al., 1997; Wallace et al., 2012). Opouawe Bank is one of the thrust-faulted and folded anticlinal ridges parallel to the margin and characterized by compression (Fig. 3.4), which leads to the tectonic build-up of the anticline. Thus, the principal compressive stress must be perpendicular to the ridge axis in NNW-SSE direction.

Formation of anticlinal ridges of limited lateral extent also leads to secondary longitudinal extension of the ridge due to gravitational forces and the flexure of the ridge (e.g. López et al., 2010, Riedel et al., 2016). Weinberger (2006) showed for Hydrate Ridge, that in the upper 200 – 400 mbsf local forces control the stress state. They infer that the topographic expression of the anticline structure controls the local stress state within the ridge, which drives extension. At Opouawe Bank the direction of longitudinal extension, is oriented along the ridge axis in ENE-WSW, as extension and folding in the central uplift zone will simply be compensated in the direction of the least compressive stress. Hence, the origin of the elongated gas migration structures in NNW-SSE direction is most likely the consequence of fractures developing as structures related to the secondary extension of the ridge and therefore parallel to the most compressive stress.

Furthermore, normal faulting at the SW flank of Opouawe Bank (Fig. 3.3), gullies at the SE flank, and translational landslide scars at the NW flank (Law et al., 2010) show that gravitational forces act on the Ridge. This is similar to Hydrate Ridge (Weinberger et al., 2006), where the ridge topography leads to gravitational collapse of its top with similar landforms such as sediment slumping and normal faulting on its eastern flank.

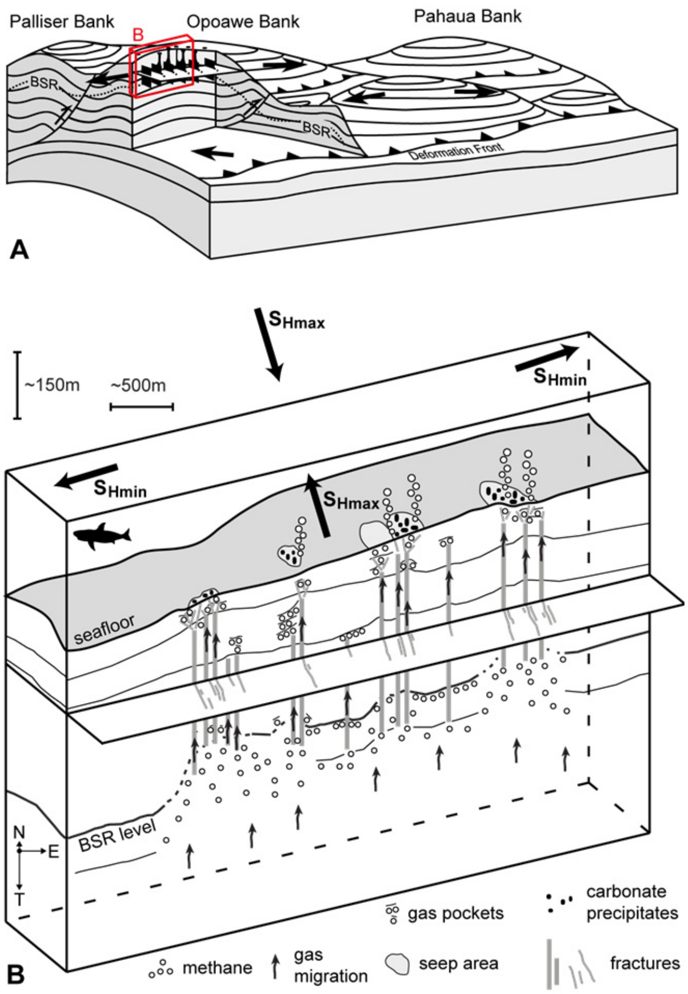

Figure 3.4 The sketches (not to scale) display Opouawe Bank in relation to the regional tectonic regime (A) and the stress regime of the ridge (B). The elongated gas migration structures are the result of an anisotropic stress regime within the anticlinal ridge. The thrust-faulted and folded accretionary ridge (A) is characterized by compressive (S_{Hmax}, - direction of maximum horizontal compression) build-up of the anticline. As a consequence of the gravitational forces, secondary longitudinal extension (S_{Hmin}, - direction of least horizontal compression) perpendicular to the direction of maximum horizontal compression takes place. The formation of gas migration structures along extensional fractures is possible through the difference between the compressive and extensional stress. In the shallow sediment declining differential stress, results in a transition to pipe-like structures.

The formation of gas migration structures along extensional fractures is possible through the difference between the compressive and extensional stress. Declining differential stress in the shallow sediment and reduction in vertical stress (overburden), results in a transition to pipe-like structures and buoyancy-driven gas migration dominates, as described below.

3.6.3. Shallow focusing of fluid flow conduits

At approximately 75 - 100 mbsf, the gas migration structures change from the elongated into more focused pipe-like pathways that culminate in sub-circular seep structures (Klaucke et al., 2010, Dumke et al., 2014). As this change in geometry does not take place at a specific horizon, we conclude that it must be controlled by other processes that depend on depth beneath the seafloor.

It seems that above 75 mbsf the buoyancy force of free gas become relatively more important than the tectonic and gravitational stresses. Hence, above this depth the process of gas accumulation, overpressure build-up, doming, and eventually gas break-through forms common circular pipe structures (Koch et al., 2015). At shallow depth the difference between hydrostatic and lithostatic pressure of the overlying sediment column decreases. The lowering of the differential stress is probably the dominant mechanism behind the focusing of flow into sub-circular features. Thus, the upward pressure driven by buoyancy can exceed the vertical pressure without the need for large-scale structures like those seen at greater depths and buoyancy-driven gas migration dominates.

Extensional fractures connect the sites North Tower and South Tower at depth, and also the Takahe and Takapu seeps appear to have common migration pathways in the lower part of the GHSZ. From this we conclude that these elongated pathways can split into several gas migration structures in the shallow sediments.

3.6.4. Seep structures at the Hikurangi margin

Numerous active cold vents occur on the crests of major accretionary ridges at the Hikurangi subduction margin (Greinert et al., 2010; Barnes et al., 2010). At Omakere Ridge different fluid migration systems exist, affected by shear and compression in a complex deformation regime (Plaza-Faverola et al., 2014). Plaza-Faverola (2014) identified four gas migration systems, with two linked to seafloor seepage. One is described as closely spaced parallel conduits with elliptical extents that they referred to as chimneys. These structures are similar to a certain degree to the observed structures on Opouawe Bank, but the conjugate nature of the structures is inferred as conjugate Riedel shear zones. The structures beneath Opouawe Bank do not have a conjugate set, just one orientation, which suggests that they are extensional fractures. The second cold seep system comprises ridge parallel extensional faults or thrusts and is associated with flexural extension of the ridge crest, whereas at Opouawe Bank extension fractures are observed perpendicular to the strike of the ridge. Extension structures perpendicular to the strike of accretionary ridges (although without associated gas migration) are observed at the northern Cascadia margin, where parallel normal faults resulting from longitudinal flexure are associated to form in the direction of the least compressive stress (López et al., 2010).

Rock Garden is influenced by uplift and extension of the ridge and Crutchley et al. (2010) showed that gas migration is connected to structural deformation, sedimentary fabrics and the gas hydrate phase boundary. The different seep sites are charged either through faults and chimneys, or along the underside of the BGHS, or along highly permeable layers that pass through the GHSZ. Generally, gas migration on Rock Garden takes place along a northwest-dipping fabric.

The wide variety of structural styles found at Opouawe Bank, Omakere Ridge and Rock Garden demonstrates the variability of geological processes within a single subduction zone. Our observation of elongated fluid migration structures is rare also in a global context, but as little high-resolution 3D seismic data exist for active subduction zones it is possible that such structures are more common than thought.

3.7. Conclusions

The analysis of gas migration pathways through Opouawe Bank shows that their geometry varies with depth including the unusual observation of elongated, parallel structures below 75 - 100 mbsf. Considering likely stress changes we conclude that a transition to pipe-like structures results from the declining differential stress and that buoyancy-driven gas migration dominates the upper sediments.

On a regional scale the stress regime is controlled by oblique subduction of the Pacific Plate underneath the Australian Plate. The build-up of anticlinal structures, such as Opouawe Bank is a consequence of the compressive setting and secondary longitudinal extension. On a local scale the stress regime within the anticlinal structure is influenced by the topography of the ridge itself. Extensional fractures are the consequence of the anisotropic stress regime within the anticlinal ridge. The great difference between the compressive stress caused by subduction and extensional stress caused by bending and gravitational forces facilitates the development of gas migration structures along extensional fractures.

3.8. References

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4. Gas-controlled seafloor doming

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4.1. Abstract

The upward migration of gas through marine sediments typically manifests itself as gas chimneys or pipes in seismic images and can lead to the formation of cold seeps. Gas seepage is often linked to morphological features like seabed domes, pockmarks and carbonate build-ups. In this context sediment doming is discussed to be a precursor of pockmark formation. Here, we present parametric echosounder, sidescan sonar and 2D seismic data from Opouawe Bank, offshore New Zealand, providing field evidence for sediment doming. Geomechanical quantification of the stresses required for doming show that the calculated gas column heights are geologically feasible and consistent with the observed geophysical data. The progression from channeled gas flow to gas trapping results in overpressure build-up in the shallow sediment. Our results suggest that by breaching of domed seafloor sediments a new seep site can develop, but contrary to ongoing discussion this does not necessarily lead to the formation of pockmarks.

4.2. Introduction

The processes that control gas migration in marine sediments and seepage are important for understanding carbon transport, especially in subduction zones, the effects of fluid migration on seafloor stability, and the occurrence of hydrocarbon resources. Migration of natural gases is frequently associated with seismically detected gas chimneys or pipes (Cartwright et al., 2007), which are often linked to seafloor features, like seabed domes, carbonate precipitates and seep fauna (Hovland and Judd, 1988). Seep evolution from the formation of bacterial mats to the development of carbonate crusts has been proposed to cause self-sealing of natural marine seeps (Hovland, 2002). Thus, seep development can impact gas migration pathways. Seabed domes have been reported from different places around the world, and have been interpreted as the result of focused fluid migration reaching the shallow sub-seafloor forming gas hydrate pingoes (Chapman et al., 2004; Hovland and Svensen, 2006) or gas domes (Hasiotis et al., 1996; Lee and Chough, 2002).

Dome structures are widely discussed to be an initial stage of pockmark formation (Lee and Chough, 2002; Judd and Hovland, 2007; Barry et al., 2012). Pingoes develop through gas hydrate expansion in the shallow subsurface, forming lenses of gas hydrate (Hovland and Svensen, 2006). Gas domes usually appear as minor topographic highs with diameters ranging from 10 to 1000 m and have been interpreted as the result of gas accumulation in near-seabed sediments (Hovland and Judd, 1988).

Opouawe Bank, offshore New Zealand, is a good site to study the processes leading to seafloor uplift. It is an anticlinal accretionary ridge (Fig. 4.1) situated in 900–1100 m water depth (Barnes et al., 2010). The sediment doming areas and seep sites investigated in this study lie inside the gas hydrate stability zone (GHSZ) (Krabbenhoeft et al., 2013). Law et al. (2010) and Krabbenhoeft et al. (2013) described the general mechanism of gas migration as structurally controlled. The most recent sediments on the ridge top are hemipelagic mud and turbidity current overspill deposits (Lewis et al., 1998). Evidence for seepage has been found in the form of gas flares in echosounder data (Greinert et al., 2010; Law et al., 2010), cold vent fauna and extended carbonate precipitates (Klaucke et al., 2010; Liebetrau et al., 2010) on the seabed, and small gas hydrate pieces in the surface sediments (Schwalenberg et al., 2010; Bialas, 2011).

Barry et al. (2012) have applied elastic thin-plate mechanics to analyze the results of laboratory experiments in which seafloor sediments were deflected. Thin-plate theory requires a wide plate width and a small deflection in comparison to the plate thickness and is therefore applicable to natural gas domes in soft cohesive sediment. In this study we extend the approach of Barry et al. (2012) by including the effects of buoyancy of the underlying gas column and comparing the resulting stresses to those required for thin-plate deformation. Comparing these results to high-resolution geophysical data from Opouawe Bank we show that thin-plate bending due to ascending gas is a viable mechanism for sediment doming and we discuss the implications of this process.

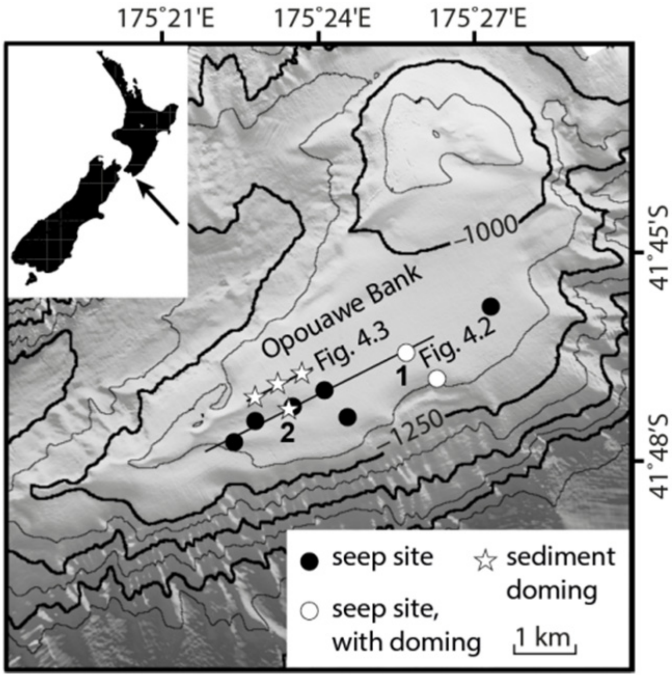

Figure 4.1: Bathymetric map of Opouawe Bank, offshore New Zealand, displaying seep sites and sediment doming areas. Dome geometries used for calculations are taken from sites 1 (Takahe) and 2 (Pukeko).

4.3. Data and methodes

4.3.1. Geophysical data

Parametric echosounder profiles (Parasound; Figures 4.2 and 4.3) were acquired over an area of 3 x 8 km with an average line spacing of 50 m, providing a unique 2.5D data set and single 2D profiles, both at high lateral resolution. The sub-seafloor was imaged with the secondary low frequency component of the system at 4 kHz, which gave penetration of up to 100 m with a vertical resolution at decimeter scale. The water column was imaged with the primary high frequency component at 18 kHz to identify emission of free gas.

The 2D multi-channel seismic (MCS) (Figs. 4.2 and 4.4) data were recorded using a 200 m long streamer with 128 channels and a group distance of 1.5 m. A single GI-gun with a volume of 210 cubic inches was operated in harmonic mode at a shot interval of 5 s. The MCS data have a useful frequency band between 40 and 300 Hz.

Seismic Unix was used for data processing including time migration (Cohen and Stockwell, 2010). Kingdom Suite (IHS) and Fledermaus (QPS) were used for interpretation and visualization of the data. Both Parasound and MCS data were combined to analyze the different manifestations of sub-seafloor structures at different frequencies and to investigate shallow gas migration pathways in the upper 100 m of sediment. Additionally, MCS data and sidescan sonar data were combined (Fig. 4.4, Dumke et al. 2014) to compare the seafloor surface expression of the seep sites with the underlying subsurface structures.

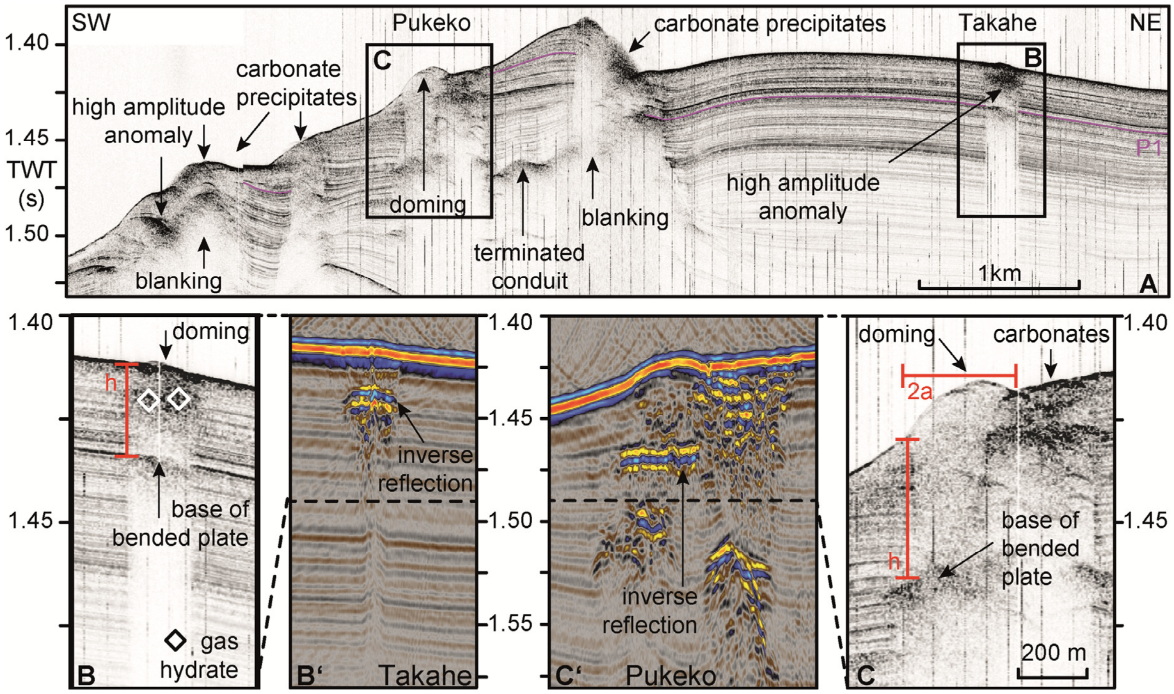

Figure 4.2: Parasound profile (A) and magnifications (B, C) with sections of coincident multichannel seismic (MCS) profile (B′, C′). Distinctive high-amplitude horizon P1 is marked in purple. Marked in red are diameter 2a and plate thickness h of dome structures. For Takahe (B), diameter is not marked, as profile crosses only marginally through the structure. Base of bent plate (B, C) appears as inverse reflection in MCS (B′,C′).

4.3.2. Pressure and gas column height

Thin-plate mechanics have been applied to natural seabed domes to quantify the pressure beneath a confining layer of sediments of known dimensions (Barry et al., 2012). This approach has been used to describe doming of soft surface sediments (Boudreau, 2012). We extend this approach by considering the gravitational forces acting on the thin plate. The accumulation of gas leads to a system dependent on the buoyancy force:

$$\Delta P = h_g \, \mathrm{g} \left(\rho_w - \rho_g\right), \tag{4.1}$$

where h_g is gas column height, g is acceleration due to gravity, and $\rho_w - \rho_g$ is density contrast between water and gas.

As ΔP in this case is the sum of the pressure accounting for the sediment plate deformation (P_{def}) and the lithostatic pressure (P_{lit}) of the sediment plate $\rho_s \mathrm{g} h$ (ρ_s is sediment bulk density, h is plate thickness), the gas column height required for providing the necessary stress can be calculated as follows:

$$h_g = \frac{\left(P_{def} + P_{lit}\right)}{\mathrm{g}\left(\rho_w - \rho_g\right)}. \tag{4.2}$$

According to Ugural (1999) and Barry et al. (2012) the pressure needed to form a dome-geometry (P_{def}), is

$$P_{def} = \frac{8}{3}\frac{E}{1-\nu}\frac{h w_{max}}{a^4}\left(\frac{2h^2}{1+\nu} + {w_{max}}^2\right) \tag{4.3}$$

where ν is Poisson's ratio, w_{max} is the maximum vertical displacement, and a is plate radius. Young's modulus E has not been measured on Opouawe Bank and therefore has to be estimated from the literature. The compilation of Barry et al. (2012) shows that E values range from 140 kPa for coastal sediments (Johnson et al., 2002) to 840 – 3,000 kPa for fine grained marine surface sediments (Lavoie et al., 1996; Wilkens and Richardson, 1998). These values were derived for sediments in shallow water depths. Because the sedimentological setting is different at Opouawe Bank, we applied a wider parameter range for E. As the recent sediments are silty clays (Bialas et al., 2007; Bialas, 2011) in water depths of around 1050 m, we approximated E for silty clay on continental shelves and slopes with a value of 350,000 kPa, after Hamilton (1971). We used a range from 140 kPa to 350,000 kPa to account for the uncertainty regarding Young's modulus. We calculate the gas column height required to form the dome-geometry of the structure at Takahe in Figure 4.2B (called case 1 in the following) and Pukeko in Figure 4.2C (case 2) with a Poisson's ratio of ν = 0.5 (Barry et al., 2012), a porosity of φ = 0.5 based on gravity cores collected along the Hikurangi margin, a grain density of 2500 kg/m^3 for the clayey sediments and 1030 kg/m^3 for the porewater. The geometry of the bent plate for case 1 is h = 20 m (0.026 s tow-way traveltime [TWT] at 1500 m/s), a = 150 m and w_{max} = 1.5 m, and for case 2 it is h = 30 m (0.04 s TWT at 1500 m/s), a = 165 m and w_{max} = 6.5 m. For comparison to other geological settings, the gas column heights have been calculated for a range of plate densities, depending on the porosity (Fig. 4.5).

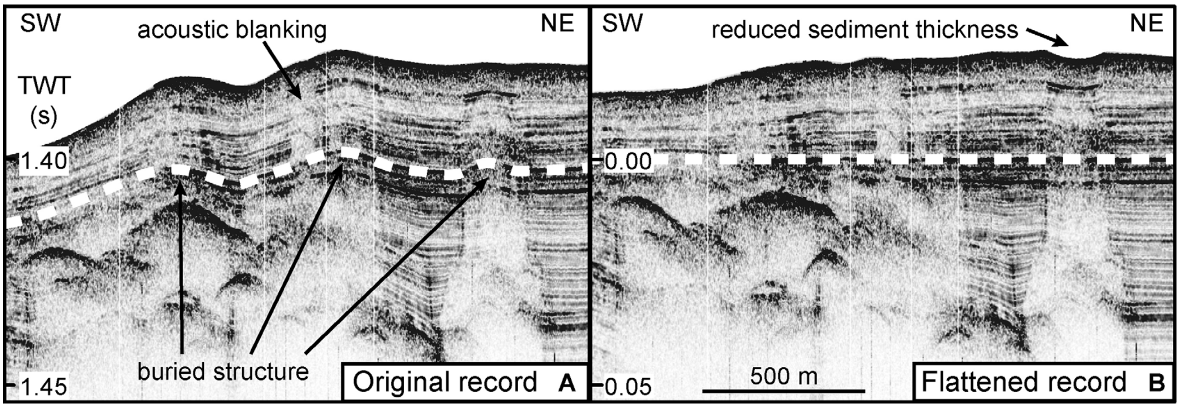

Figure 4.3: A: Buried dome structures. B: Same section with flattening of domed horizon, with the time scale relative to the flattened horizon in two-way traveltime (TWT, s). Flattened record displays less sedimentation on top of buried dome structures.

4.4. Results

Domes observed at Opouawe Bank originate at the top of seismic pipes and deflections are 1 – 2% of the dome diameters. The magnitude of vertical doming is ~ 2 – 6 m above the surrounding seafloor. Up-bending of the sediment is observed at the seafloor unaffected by seepage (Fig. 4.2C), at seep sites without carbonate precipitates (Fig. 4.2B), and for buried structures (Fig. 4.3). In the shallow subsurface, sediment interfaces can be mapped continuously over the entire survey area. Amplitude anomalies within the gas migration structures occur preferentially in continuation of high-amplitude horizons, partly associated with a local reduction of continuity and high-amplitude anomalies with signal attenuation or acoustic blanking underneath. One prominent example is horizon P1 (Fig. 4.2), has distinct high-amplitude anomalies at all structures and which changes polarity within some structures (Fig. 4.2B, 4.4C). Active gas emission was observed at all seep sites during the time of surveying. Seep sites with carbonate precipitates at the seafloor (Klaucke et al., 2010) are not associated with doming.

Seep structures with slightly domed sediment (Fig. 4.2B) and bacterial mats at the seafloor (Greinert et al., 2010) reveal a strong reverse polarity reflection in the MCS data 20 mbsf at horizon P1, with patchy less pronounced amplitude anomalies underneath. The Parasound data additionally display a bright diffuse reflection between horizon P1 and the seafloor. Gravity cores taken at the intersection of this reflection with the seafloor contained a few small gas hydrate pieces (Schwalenberg et al., 2010; Bialas, 2011).

Figure 4.2C displays two juxtaposed gas migration structures characterized by different reflection patterns. Although the Parasound data indicate that both structures extend to the seafloor, the observed carbonate blocks from the sidescan sonar image occur only over the eastern structure (Fig. 4.4) (Dumke et al., 2014). At the western structure the seafloor domes upwards. Notably, the two flares observed at this site occurred between the two structures and on the eastern rim of the eastern structure.

For the buried doming structures (Fig. 4.3), an artificial flattening of the horizon on top of the high-amplitude anomalies reveals less sedimentation above the domes. We interpret this as evidence of older doming structures buried by hemipelagic sediment. Narrow zones of blanking appear to reach the seafloor above the buried domes, indicating the presence of gas but we do not observe gas flares and there is no expression of seepage in the sidescan sonar data (Dumke et al., 2014).

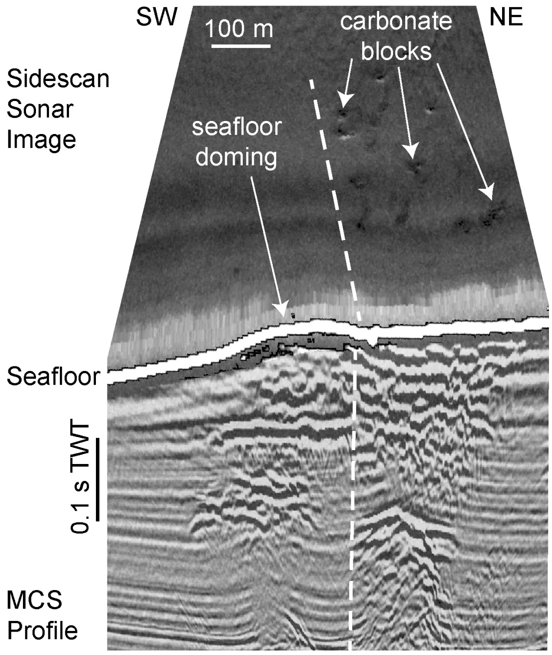

Figure 4.4: Combined view of sidescan sonar and multichannel seismic (MCS) data for Pukeko seep site (offshore New Zealand). Seafloor backscatter map of sidescan sonar shows distribution of seep carbonates which cause high backscatter (black) due to high impedance contrast and surface roughness. MCS data show that subsurface seep structure consists of two parts, with seafloor doming over western part and carbonate precipitation over eastern part.

The pressures affecting the sediment dome system and the resulting gas column heights (Fig. 4.5B) were calculated for the dome geometries of the sites Takahe (case 1, Fig. 4.2B) and Pukeko (case 2, Fig. 4.2C). For Case 1 the possible gas column height lies within the range of 37.6 m (ΔP ~346.3 kPa) to 43.9 m (ΔP ~404.7 kPa) and for case 2 between 56.4 m (ΔP ~519.7 kPa) and 121.5 m (ΔP ~1120.8 kPa). For case 2 the plate thickness cannot be determined precisely and could be thinner than estimated and thus, closer to the results of case 1. The calculations demonstrate that the plate thickness and Young's modulus have a major impact on the calculated height of the gas column. The porosity of 0.5 on Opouawe Bank is relatively low and a higher porosity of the sediment, as found in various marine settings, would cause a smaller bulk density resulting in a reduced gas column height.

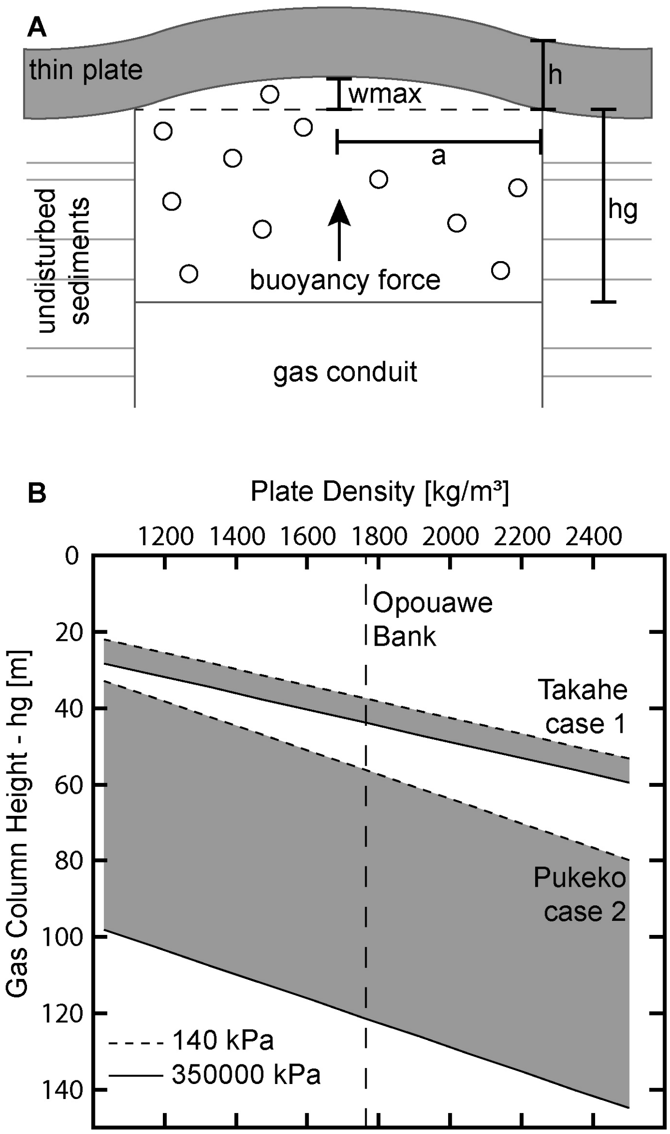

Figure 4.5: Thin-plate bending in response to buoyancy force of accumulating gas (illustrated by circles), with plate thickness h, radius a, maximum vertical displacement wmax, and gas column height hg. B: Diagram of gas column height versus plate density for Takahe (h = 20 m, a = 150 m, wmax = 1.5 m) and Pukeko (h = 30 m, a = 165 m, wmax = 6.5 m) for range of 140–350,000 kPa for Young's modulus. Dashed line indicates sediment bulk density on Opouawe Bank.

4.5. Discussion

4.5.1. Sediment doming and seepage

Upward gas migration through the GHSZ at Opouawe Bank is structurally controlled (Krabbenhoeft et al., 2013) and continues until the shallow subsurface is reached. Here, gas accumulation leads to overpressure build-up, doming and eventually gas break-through, which might occur in dense clusters of individual gas bubble pathways (Haeckel et al., 2007).

Indications for gas accumulation are found mainly underneath sedimentary horizons with strong reflectivity. Truncated reflections indicate that the gas trapping horizons may not be continuous beneath the active seep sites. Our interpretation is that gas accumulates below less-permeable horizons and causes sediment doming due to buoyancy. The high-amplitude reversed-polarity reflections in the MCS data and the acoustic blanking in the Parasound data strongly suggest that the sedimentary layers underneath the domes are gas charged and an increase in the gas concentration is to be expected. As discussed above the required gas columns for sites off New Zealand are between 35 m and 120 m depending on geometry and Young's modulus. Comparing these values to the seismic observations (Figs. 4.2B and 4.2C) we find that there are high-amplitude anomalies extending more than 100 ms TWT below the up-bent plates. Although it is not possible to image a distinct base of the gas columns with the MSC data, the gas column heights of several tens of meters are consistent with our seismic observations of >100 m high gas columns and seem geologically plausible.

It is unlikely that the existence of the dome structures is caused by gas hydrate formation. Only a few gas hydrate pieces were recovered at two seep structures in the study area. Doming by gas hydrate formation would require high hydrate saturations and there is no evidence in the Parasound data (Fig. 4.2) that hydrate saturation increases significantly below the depth for which we have control from sediment cores. Furthermore, based on sidescan sonar, video (Klaucke at al., 2010; Dumke et al., 2014), and Parasound data we can rule out dome formation by carbonate build-ups.

4.5.2. Buried domes

The buried domes (Fig. 4.3) indicate that gas migration at these structures halted at some point in the past. These structures potentially resulted from the same mechanism that leads to sediment doming at the seafloor, as discussed above, but the high-amplitude, up-bent reflections could also represent buried carbonates. Above the buried domes zones of seismic blanking are present. We conclude that a reactivation of gas migration underneath the domes leads to breaching of trapping layers and thereby to the establishment of new pathways. If the buried dome structures have indeed formed by past seafloor doming, it would imply that the dome morphology can sustain burial. This may be due to elevated pore pressures, but could also result from permanent structural deformation of the surface sediments during formation of the domes.

4.5.3. Up-bending versus pockmark formation

Dome structures are widely discussed to be an initial stage of pockmark formation (Lee and Chough, 2002; Judd and Hovland, 2007; Barry et al., 2012). Hovland and Judd (1988) developed a conceptual model for the formation of pockmarks. In this model doming is a result of inflating the sediment column as gas enters from beneath and represents a zone of tension where the sediments are stretched, which is accompanied by fracturing. A pockmark would form as a result of pressure drop when a connection to the seafloor is made with a violent outburst of escaping gas. In contrast to this model, we do not observe pockmarks. We rather suggest that domes evolve gradually by fracturing of gas trapping horizons and the establishment of preferential pathways for gas migration. Over time, doming and break-through will allow gas to migrate all the way toward the seafloor as observed at the Takahe seep (Fig. 4.2B). Seafloor doming becomes the last stage before active gas emission finally forms a new cold seep not necessarily associated to a pockmark.

4.6. Conclusions

High-resolution sub-bottom profiling data constrain the nature of shallow gas migration beneath Opouawe Bank. Clear differences in individual gas migration structures indicate a progression through different evolutionary stages, which range from channeled gas flow and associated seismic blanking, to gas trapping beneath relatively low-permeable horizons, and finally overpressure accumulation and doming. Calculating the height of the gas column necessary to create different dome geometries, shows that doming due to gas accumulation is feasible and consistent with field observations.

The process of gas accumulation beneath particular layers and subsequent doming appears to be a precursory process in the development of methane seep sites on Opouawe Bank and might be a common characteristic for gas seeps in general. The well-stratified, sub-horizontal strata that exist beneath Opouawe Bank provide favorable conditions for this type of seep development, because shallow sub-vertical gas migration is forced to traverse sedimentary layers in the absence of faults that might otherwise have provided more efficient gas migration pathways. Thus, the gas has to generate its own migration pathways through the progressive process of doming and breaking through the strata. The data from offshore New Zealand document that upward migration of gas and associated shallow sediment doming does not have to be associated with seafloor pockmarks and that models in which fluid migration and associated doming in soft sediments necessarily culminates in pockmark formation are not applicable everywhere.

4.7. Acknowledgements

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5. Conclusions and outlook

5.1. Summary of the key results

This work presents new insights into fluid flow systems on the basis of geophysical and geochemical investigations at Opouawe Bank, on the Hikurangi margin off New Zealand. Three case studies highlight the different aspects of gas migration from the gas source to the seabed.

Chapter 2

Gas migration through Opouawe Bank at the Hikurangi margin, offshore New Zealand

This study addressed the source and plumbing system of gas seeps on the Hikurangi margin. The combination of 2D MCS data, seismic velocity analysis, geochemical, and isotopic porewater compositions augments our ability to investigate gas migration processes beneath Opouawe Bank.

The MCS data and velocity analysis showed that gas migration is influenced by focusing through the anticlinal ridge structure. Gas accumulates at the core of the anticline beneath the GHSZ, as indicated by a LVZ in about 320 mbsf. At the center of the ridge, the BSR is mostly absent, the LVZ extends upwards into the GHSZ. Some gas migration takes place along inclined sedimentary strata, suggesting the absence of a sharp impedance contrast across the BGHSZ. The structurally controlled gas migration along vertical conduits from the LVZ through Opouawe Bank is characterized by focusing at the base and within the GHSZ.

Gas seepage is driven by upward migration of microbial gas. This is demonstrated by the absence of mineral dewatering reaction signals, in the porewater isotopic signature (of oxygen and hydrogen) and chlorine (Cl) concentrations. In addition the porewaters are not enriched in boron and lithium. The gas primarily consists of methane, without a thermogenic imprint and is generated by the degradation of organic matter via microbial CO_2 reduction. Bubble-induced irrigation-type porewater profiles indicate that methane transport is dominated by gas migration. The methane largely originates below the GHSZ and the maximum source depth is 1,500–2,100 m below seafloor.

Chapter 3

Elongated fluid flow structures: Stress control on gas migration in Opouawe Bank, New Zealand

This study revealed the exceptional nature of focused gas migration within the GHSZ at Opouawe Bank. The 3D seismic data analysis showed parallel and horizontally elongated structures. They exhibit a larger extent perpendicular to the ridge-axis, and a transition to pipe-like structures in the shallow subsurface from about 75 – 100 mbsf up to the seafloor. This is, however, different from typical manifestations of focused vertical gas migration structures, such as pipes and chimneys on seismic images.

The elongation of gas migration structures is the result of an anisotropic stress regime within the anticlinal ridge. The local stress regime of the thrust-faulted and folded accretionary ridge is characterized by compressive build-up of the anticline. As a consequence of the gravitational forces, secondary longitudinal extension perpendicular to the compressive stress takes place. The difference between the compressive and extensional stress facilitates the development of gas migration structures along extensional fractures. Declining differential stress in the shallow sediment and reduction in vertical stress (overburden), results in a transition to pipe-like structures and buoyancy-driven gas migration dominates.

Chapter 4

Gas-controlled seafloor doming

The investigation of shallow gas migration structures on Opouawe Bank provides advancement on the current theory of seabed domes and gas migration in marine sediments. Dome structures are considered to be an initial stage of pockmark formation. However, the analysis of 2.5D Parasound, 2D MCS, and side-scan sonar data from Opouawe Bank shows that contrary to ongoing discussion domes can evolve into seep sites and do not necessarily evolve into pockmarks. Clear differences in individual gas migration structures indicate a progression through different evolutionary stages of seep formation, which range from channeled gas flow, to gas trapping beneath relatively low-permeable horizons, to overpressure accumulation and doming. By breaching the domed seafloor sediment a new seep site can develop.

The Parasound data display doming structures in great detail; this allowed the synthesis of field data with a model for dome formation. Thin plate mechanics have been applied to natural seabed domes to quantify the pressure causing the deflection of the dome sediment (Barry et al., 2012). Extending this theoretical approach by considering the gravitational forces acting on the thin plate, for the first time, the gas column height underneath the domed structures could be quantified.

5.2. Implications

This thesis was part of the NEMESYS project on the Hikurangi margin off the east coast of New Zealand’s North Island, with the focus on the geophysical data at Opouawe Bank. The general aim of the project was to improve the knowledge of seep sites and their variability at active continental margins, and review present models of the build-up of seeps and their feeder structures with respect to the tectonic regime.

Comparing fluid migration structures at anticlinal ridges along the Hikurangi margin, a wide variety of structural styles is observed, with some similarities to Opouawe Bank (e.g. Crutchley et al., 2010; Plaza-Faverola et al., 2014). At Opouawe Bank the spatial analysis suggests that gas migrates along extensional fractures. The elongated fluid migration structures are rare in a global context, but as little high-resolution 3D seismic data exist for active subduction zones it is possible that such structures are more common than thought. For instance, parallel extension structures perpendicular to the strike of accretionary ridges have been observed at the northern Cascadia margin (López et al., 2010; Riedel et al., 2016), although without associated gas migration. An interesting study prospect would be a more detailed analysis of the spatial structure of gas migration pathways at accretionary ridges, with new 3D seismic data or re-interpretation of existing data sets. Especially the reevaluation of existing data might provide important insight. This is demonstrated by the consecutive studies at Opouawe Bank, where already indications for elongated structure were observed by Krabbenhoeft et al. (2013), as an offset of chimney structures with respect to the seeps location on 2D seismic profiles.

The gas migration structures convert to pipe-like structures in the shallow sediment and gas accumulation, overpressure build-up, doming, and eventually gas break-through is dominant process. Gas-controlled seafloor doming is of interest to a wide range of disciplines including the broad categories of sediment mechanics, stability, and failure; gas and oil, and gas hydrates. The phenomenon is important for fluid migration and the formation (or not) of fluid escape structures at the seabed in cohesive soils (clays to silty clays). A better understanding of gas migration and the stability of seabed doming in areas of gas and oil exploration is relevant to improve the safety of operations and provides an additional tool to understand the mechanics of upward migrating gas through marine sediments. Unfortunately no site-specific information is available for Young's modulus. Consequently, a large parameter space was used. To better constrain gas column heights required for such doming structures, geotechnical and soil data would be required, which is something marine geoscience often lacks.

As part of the multidisciplinary NEMESYS project, marine controlled source electromagnetic (CSEM) data were collected on Opouawe Bank. They indicate the presence of anomalously high resistivities (3-100 Ωm) beneath several seep sites (Fig. 5.1). The complimentary seismic data show focused fluid flow along linear extensional fractures, which are a possible source of electrical anisotropy. The marine CSEM data cannot resolve electrical anisotropy, but is generally consistent with a simplified anisotropic model and a resistive top layer. The resistive top layer is coincident with the transformation of the gas migration structures to pip-like, circular structures. Katrin Schwalenberg (BGR, Germany) is preparing a paper on the geophysical signatures of methane seepage and gas hydrate formation at Opouawe Bank, by combining CSEM and seismic data.

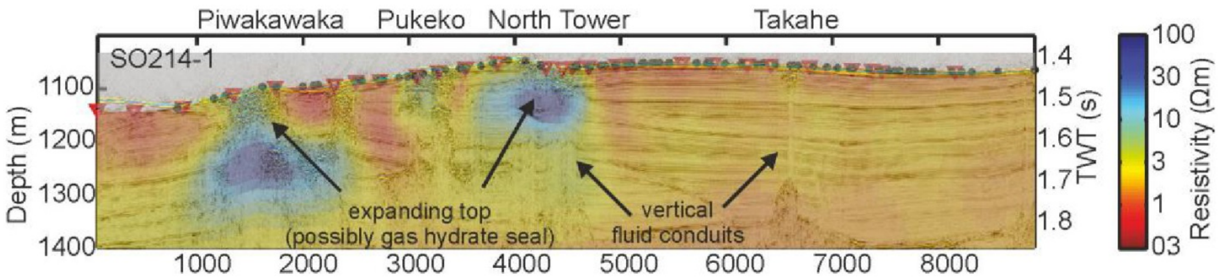

Figure 5.1: Overlay of a 2D electrical resistivity model and the coincident seismic reflection profile along Opouawe Bank (from Rippe et al., 2014).

Far-reaching implications of the study encompass the entire subduction zone system, including megathrust earthquake cycles and carbon-flux issues. As one of the few studies utilizing high resolution 3D seismic imaging, this work showed for the first time the ability to map the horizontal stress regime. Although within the compressive field of a subduction accretionary prism, extensional features allow the delineation of orientation in maximum and minimum horizontal compressive stresses. The magnitude of these stresses relative to the overburden stress (lithostatic) can be used to determine the behavior of the underlying megathrust fault (e.g. Lin et al., 2013, 2015; Malinverno et al., 2016). A key question here is at what tectonic stage the Hikurangi margin is within the megathrust cycle. The Cascadia margin is at a late stage after the last megathrust quake in 1700 and studies off New Zealand show that the last megathrust quake off the coast of Wairarapa (west of Opouawe Bank) may have occurred 470-520 years ago (e.g. King et al., 2007), indicating a late stage in the interseismic period. Yet, other recent earthquakes in the region in 1855 (Wairarapa coast) and 1947 (further North, Gisborne earthquake may also ruptured portion of the subduction megathrust (Wallace et al., 2014). Using the Nankai accretionary margin as an example, Wang and Hu (2006) proposed that the outer prism undergoes sub-horizontal compression during a megathrust earthquake due to velocity-strengthening of the megathrust. Their model predicts that during the interseismic phase (i.e. seismogenic zone is locked), the outer prism becomes increasingly relaxed until dominated by gravity effect (overburden stress).

The large-scale implications of this study for the carbon flux encompass linkages between the tectonic stress regime and the distribution of vent sites. These vent-locations are governed by physics. Vent systems and chemosynthetic communities are "dependent" on the formation of fluid pathways. Carbon flux across the subduction zone is thus heterogeneous, but not fully random. However, while the focused flux at vent sites can be captured through the study of the stress-regime along and across the prism, pervasive or diffuse flux is not governed by these tectonic processes. A key finding of this study is that the vent systems of Opouawe Bank, Rock Garden, and Omakere Ridge are gas-dominated with little fluid/pore water migration (i.e. a “dry system”) which is in contrast to e.g. Cascadia, where ample evidence is available for fluid migration (Riedel et al., 2010). If the system is dry, questions arise to the fate of the water within the subduction zone system, e.g. whether fluid is transported further landward into the more mature accretionary prism before it migrates upward. Alternatively, a question is whether fluid that is expelled from the prism migrates along a decollement to the toe of the prism (i.e. seaward), thus escaping the sedimentary structures of the ridges (e.g. at Barbados, Bangs et al., 1999). As water in the subduction system can also influence the extent of the locked zone (e.g. Ranero et al., 2008; Worcewski et al., 2011) studies on the geochemistry of vent systems can provide constraints on the extent of the potential rupture zone.

5.3. Outlook

This work has contributed to an improved understanding of fluid flow and related processes on the Hikurangi margin. However, there are still aspects that are not yet fully understood. The following outlook presents ongoing and planned future surveys that are related to fluid migration, and the gas hydrate system, offshore New Zealand.

Gas hydrates as an energy resource

The research program into gas hydrates as an energy resource in New Zealand (funded by the New Zealand Ministry of Business, Innovation, and Employment), had a mid-term review in 2015 (after 3 years). The program has been extended for the next three years. One of the key focuses is to use geophysics (i.e. seismics) to characterize the most attractive sites of gas hydrate deposits that could be exploited in the future.

Production modelling studies are conducted by Ingo Pecher and his colleagues at the University of Auckland. They use the code "TOUGH + HYDRATE" (Moridis et al., 2014) to simulates how a gas hydrate reservoir responds to production by depressurization. The idea is to see how effective gas production is from a reservoir, given a set of boundary conditions, including permeability and hydrate saturation. Another field of active research in this program, led by NIWA (National Institute of Water and Atmospheric Research) is the relationship between gas seepage and seafloor biological communities. Furthermore, gas hydrate formation modelling will continue with Petromod, i.e. led by Karsten Kroeger from GNS (Institute of Geological and Nuclear Sciences).

MeBo drilling

Led by Katrin Huhn (MARUM Center for Marine Environmental Sciences Bremen, Germany) and Nina Kukowski (University Jena, Germany), the RV Somme SO247 cruise, as part of the SLAMZ (Slide activity on the Hikurangi margin, New Zealand) project was conducted in March and April 2016. The project used the MeBo drill rig (capable of coring up to 200m) to investigate the connection of gas hydrate dissociation and slope failure processes in two study areas, Rock Garden and the Tuaheni slide complex. The plan was to drill at Tuaheni, to sample much (if not all) of the gas hydrate stability zone, and providing the first deep lithological information about where gas hydrates form. No scientific drilling has been done before on the Hikurangi margin, only industry oil and gas exploration drilling. The second study site for MeBo was Rock Garden, to improve the understanding of the influence of both seamount subduction and gas hydrate stability on seafloor erosion at the accretionary ridge.

The campaign was complemented by gravity cores, heat flow measurements and hydro-acoustic mapping. At Rock Garden several cores, with up to 35 m penetration were taken at the same location. It revealed about 20 m clayey silt with sections of turbidite material and tephra layers and laminated clayey silt below. Practically no methane was observed in upper 20m, with an unusual concentration gradient below. The recovered cores will help to comprehend the sediments behavior during uplift and erosion on the ridge and to assess the hypothesized mechanical weakening of the ridge top.

At Tuaheni 2 holes were drilled in the landslide and 2 outside with MeBo. The lithology within the landslide contains significant amounts of coarse silt and thus, the recovery rate of the cores was non-satisfying. The original plan was to sample much of the gas hydrate stability zone, and providing the first deep lithological information about where gas hydrates form, however no gas hydrate was recovered from the slide.

IODP proposal

Ingo Pecher (University of Auckland, New Zealand) is leading an "Ancillary Project Letter" to IODP. The proposal was linked to the results of the SO247 RV Sonne MeBo drill cruise (see above) and is now approved. The objectives are to test several hypotheses concerning the cause for creeping of the Tuaheni slide complex, based on the (assumed) occurrence of gas hydrates. These include ice-like viscous behavior due to gas hydrates, overpressure transmitted into the GHSZ and subsequent sediment weakening, and slow sliding at the BGHS due to overpressure. An antithesis is also considered, in case gas hydrates are not involved, assuming repeated small-scale failure.

Gas flux and gas source for gas hydrates on the northern Hikurangi margin

This study builds on the RV Tangaroa voyage TAN1508 to geochemically assess the Hikurangi margin shallow sediment off Mahia Peninsula, east of New Zealand’s North Island (Coffin et al., 2015). Pore-water and gas geochemistry is used to predict the flux and source of gas, which supplies the gas hydrate system. One of the key targets in this project is to combine seismic images with geochemistry information to understand regions where gas flux into the hydrate layer is more pronounced than elsewhere.

Submarine landslides

The BMBF funded MAWACAAP (Quantifying the role of mass wasting in submarine canyons on active and passive margins) project led by Sebastian Krastel (University of Kiel) and Joerg Bialas (GEOMAR, Kiel) studies submarine landslide processes in the Cook Straight and Kaikoura regions (southern Hikurangi margin). One of the target areas is Palliser Bank, an accretionary ridge adjacent to Opouawe Bank, on the southern end of the Hikurangi subduction zone. On Palliser Bank there is evidence for landslide processes related to the pinch out of the BGHSZ. Thus, the objective is to explore the relationships between gas hydrates, fluid flow and slope stability in that region.

Methane seepage offshore Poverty Bay

Joshu Mountjoy (NIWA, New Zealand) is leading the work on a project looking further into shallow gas seeps discovered from multibeam data during TAN1404 cruise with RV Tangaroa, off. Multibeam and single-beam water column imaging is used to estimate the flux of methane coming out of some of those seep sites. The ultimate aim is to make a prediction of the overall gas flux out of this area into the water column.

Fluid flow processes, Fjordland margin

Marion Jegen (GEOMAR Kiel, Germany) submitted a proposal to look into fluid flow processes on the Fiordland margin that are related to the initiation of subduction on that margin. If successful, part of the work in that project will be to look at how fluids flow through the hydrate system, to better understand the deep fluid flow processes related to tectonics.

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Appendix A

A.1 Supplementary Material Chapter 2

Table A.1: Locations of TV-guided multiple cores (MUC) and gravity cores (GC) at the various seep sites and areas at the Hikurangi Margin

Area / **Seep**	**Core**	**Latitude (S)**	**Longitude (E)**	**Water depth / m**	**Core length / cm**
Omakere Ridge					
LM-9	50 GC 1	40°01.070'	177°51.585'	1059	430
	64 GC 2	40°01.067'	177°51.581'	1150	490
	244 GC 25	40°01.068'	177°51.583'	1152	477
Kaka	65 GC 3	40°02.160'	177°48.056'	1165	425
	243 GC 24	40°02.158'	177°48.049'	1167	560
Bear's Paw	93 GC 5	40°03.188'	177°49.162'	1100	170
	94 GC 6	40°03.183'	177°49.182'	1101	116
	95 GC 7	40°03.192'	177°49.174'	1100	90
	196 GC 20	40°03.196'	177°49.183'	1102	170
Rock Garden					
Faure Site	217 GC 21	40°01.138'	178°09.703'	662	134
	267 GC 27	40°01.937'	178°09.653'	662	115
Opouawe Bank					
North Tower	111 GC 8	41°46.955'	175°24.184'	1061	254
	124 MUC 12	41°46.927'	175°23.988'	1048	17
	125 GC 10	41°46.932'	175°23.973'	1041	150
	152 GC 12	41°46.956'	175°24.182'	1065	490
	273 MUC 40	41°46.962'	175°24.268'	1059	22
	285 GC 37	41°46.956'	175°24.189'	1061	286
	287 GC 39	41°46.961'	175°24.233'	1059	400
South Tower	295 GC 40	41°47.595'	175°24.247'	1081	374
	296 GC 41	41°47.377'	175°24.473'	1056	360
	302 GC 42	41°47.327'	175°24.541'	1056	150
Tui	112 GC 9	41°43.299'	175°27.110'	815	47
Takahe (*GH)	304 GC 43	41°46.340'	175°25.580'	1054	280
References					
Omakere Ridge	79 GC 4	40°01.401'	177°48.941'	1180	540
	169 GC 15	40°01.415'	175°49.011'	1177	591
	170 GC 16	39°56.350'	177°55.350'	1155	584
Opouawe Bank	153 GC 13	41°45.444'	175°26.536'	1057	535

(*GH) = gas hydrate samples at Takahe seep were analyzed for gas composition and CH4 isotopy

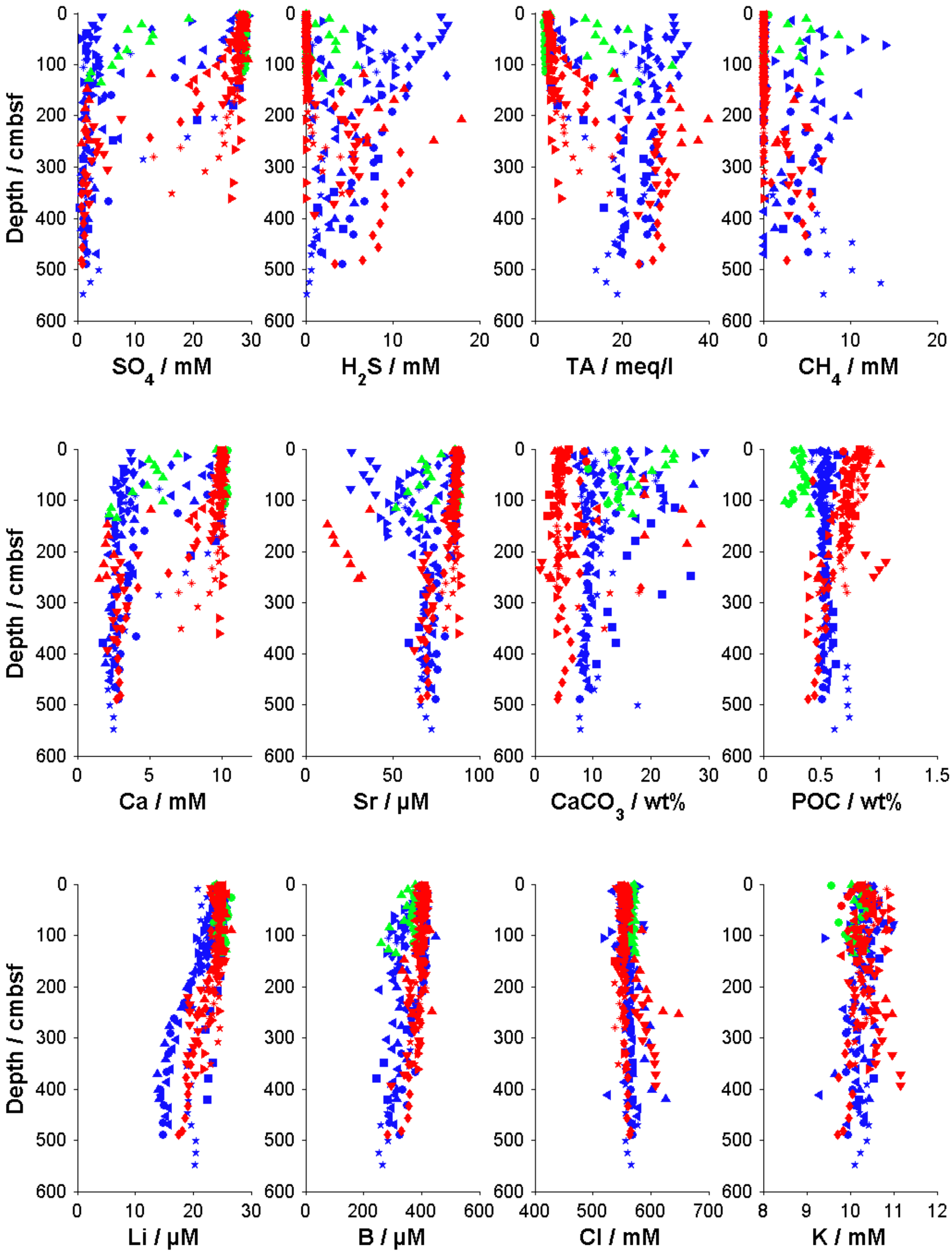

Figure A.1: Concentration-depth profiles of various dissolved porewater and solid phase species from gravity cores taken at various seep sites (different symbols) in the investigation areas (green: Rock Garden; blue: Omakere Ridge; red: Opouawe Bank) along the Hikurangi Margin (see Tab. A.1 for details).

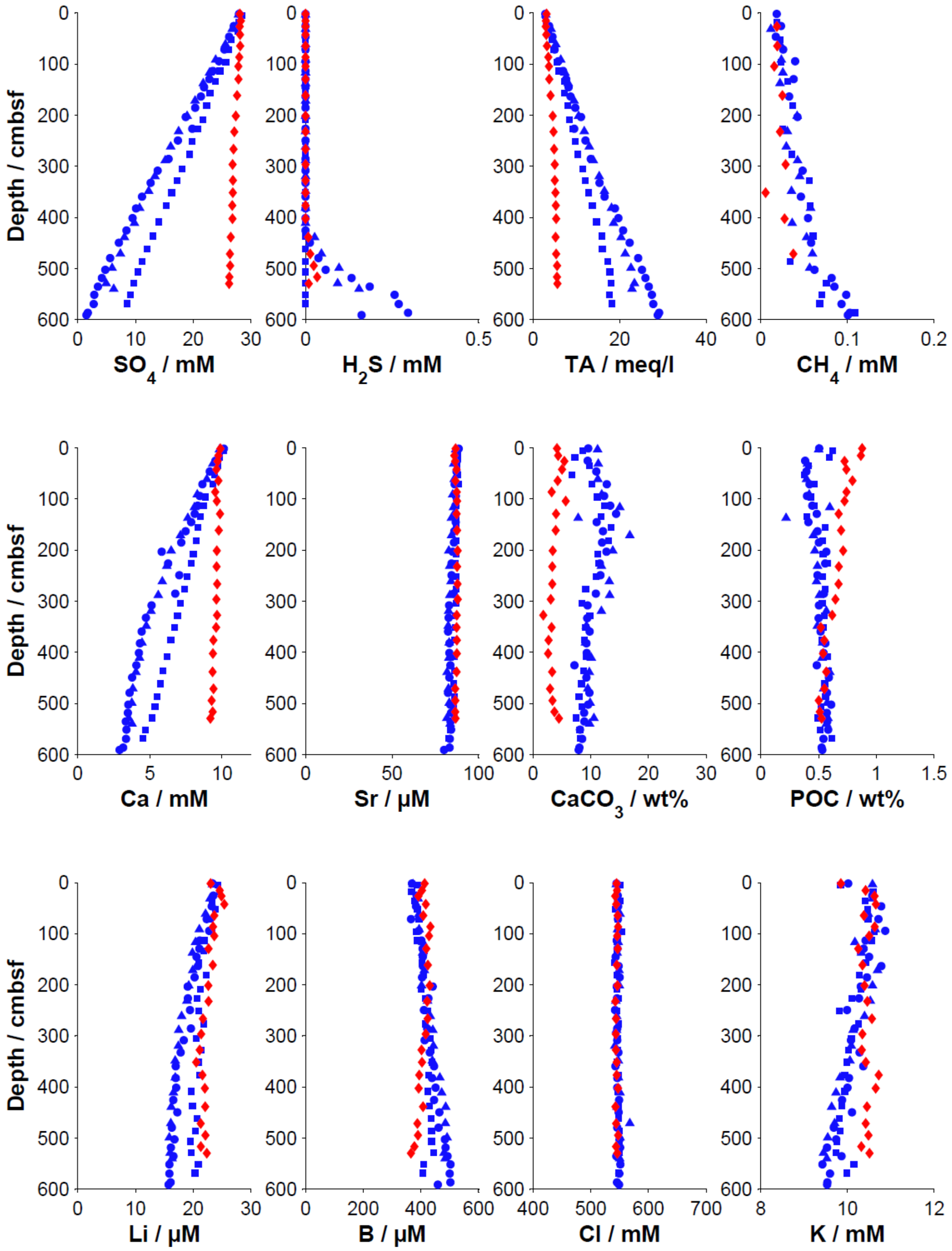

Figure A.2: Concentration-depth profiles of various dissolved porewater and solid phase species from gravity cores taken at the reference sites (different symbols) in the investigation areas (blue: Omakere Ridge; red: Opouawe Bank; see Tab. A.1 for details).

Appendix B

B.1 Curriculum Vitae

Stephanie Koch

Date of birth	6 November 1986
Place of birth	Erfurt
Citizenship	German
Permanent address	Nettelbeckstraße 2, 24105 Kiel, Germany

Education

since 02/2012	PhD student at GEOMAR and Christian-Albrechts-University, Kiel
01/2012	MSc in Geophysics
04/2009 – 01/2012	studies in Geophysics at Christian-Albrechts-University, Kiel
03/2009	BSc in Applied Geophysics
10/2005 – 03/2009	studies in Applied Geophysics at Christian-Albrechts-University, Kiel
05/2005	Abitur
08/1997 - 05/2005	Königin-Luise-Gymnasium, Erfurt, Germany

B.2 List of publications and presentations in the period of my PhD

Articles - peer-reviewed

Koch, S., Berndt, C., Bialas, J., Haeckel, M., Crutchley, G., Papenberg, C., Klaeschen, D. and Greinert, J. (2015). Gas-controlled seafloor doming. Geology, 43, 571-574, doi:10.1130/G36596.1.

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Presentations

Gross, F., Mountjoy, J., Crutchley, G., **Koch, S.**, Bialas, J., Pecher, I., Woelz, S., Dannowski, A., Carey, J., Micallef, A., Böttner, C., Huhn, K. and Krastel, S. (2016) Submarine creeping landslide deformation controlled by the presence of gas hydrates: The Tuaheni Landslide Complex, New Zealand. In: EGU General Assembly 2016, 17.-22.04.2016, Vienna, Austria

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